ADVANCING RIVER RESTORATION AND MANAGEMENT

SECOND EDITION

TOOLS IN FLUVIAL GEOMORPHOLOGY

EDITED BY

G. MATHIAS KONDOLF AND HERVÉ PIÉGAY

WILEY Blackwell

Tools in Fluvial Geomorphology

Tools in Fluvial Geomorphology

Second Edition

Edited by

G. Mathias Kondolf

University of California, Berkeley, USA

and

Hervé Piégay

CNRS, University of Lyon, France

ADVANCING RIVER RESTORATION AND MANAGEMENT

WILEY Blackwell

Registered office: John Wiley & Sons, Ltd, The Atrium, Southern Gate, Chichester, West Sussex, PO19 8SQ, UK

Editorial offices: 9600 Garsington Road, Oxford OX4 2DQ, UK
The Atrium, Southern Gate, Chichester, West Sussex PO19 8SQ, UK
111 River Street, Hoboken, NJ 07030-5774, USA

For details of our global editorial offices, for customer services and for information about how to apply for permission to reuse the copyright material in this book please see our website at www.wiley.com/wiley-blackwell.

Library of Congress Cataloging-in-Publication Data

Names: Kondolf, G. Mathias, editor of compilation. | Piégay, Hervé, editor of compilation.
Title: Tools in fluvial geomorphology / edited by G. Mathias Kondolf and Hervé Piégay.
Description: Second edition. | Chichester, UK ; Hoboken, NJ : John Wiley & Sons, 2016. | Includes index.
Identifiers: LCCN 2016006833 | ISBN 9780470684054 (cloth)
Subjects: LCSH: Fluvial geomorphology. | Geomorphology–Instruments.
Classification: LCC GB562 .T66 2016 | DDC 551.3/5–dc23 LC record available at http://lccn.loc.gov/2016006833

A catalogue record for this book is available from the British Library.

Wiley also publishes its books in a variety of electronic formats. Some content that appears in print may not be available in electronic books.

Set in 9.5/12pt Minion by SPi Global, Chennai, India

1 2016

Contents

List of contributors

Kazutake Asahi
RiverLink Corporation
Tokyo, Japan
kazutake.asahi@river-link.co.jp

Gudrun Bornette
UMR CNRS 6249 Chronoenvironnement
Université de Bourgogne Franche-Comté
25030 Besançon, France
gudrun.bornette@univ-fcomte.fr

Anthony G. Brown
Department of Geography
University of Southampton
Southampton SO17 1BJ, UK
tony.brown@soton.ac.uk

Robert Bryant
Department of Geography
University of Sheffield
Sheffield SN102TN, UK
r.g.bryant@sheffield.ac.uk

Janine Castro
US Fish and Wildlife Service
Portland, OR 97266, USA
janine_m_castro@fws.gov

Stephen E. Darby
Department of Geography
University of Southampton
Southampton SO17 1BJ, UK
s.e.darby@soton.ac.uk

Peter W. Downs
School of Geography, Earth and Environmental Sciences
Plymouth University
Plymouth, PL4 8AA, UK
peter.downs@plymouth.ac.uk

Simon Dufour
Département de Géographie
UMR CNRS LETG Rennes COSTEL
Université Rennes 2
35000 Rennes, France
simon.dufour@univ-rennes2.fr

Thomas Dunne
Donald Bren School of Environmental Sciences and Management and
Department of Geological Sciences
University of California, Santa Barbara
Santa Barbara, CA 93106, USA
tdunne@bren.ucsb.edu

Ian Foster
School of Science and Technology
University of Northampton
Northampton NN2 6JD, UK
ian.foster@northampton.ac.uk

David Gilvear
School of Geography, Earth and Environmental Sciences
University of Plymouth
Plymouth PL48AA, UK
david.gilvear@plymouth.ac.uk

Basil Gomez
Department of Geography
University of Hawaii
Manoa, Honolulu, HI 96822, USA
basil@kbayes.com

Robert C. Grabowski
Cranfield Water Science Institute
Cranfield University
Cranfield MK43 0AL, UK
r.c.grabowski@cranfield.ac.uk

Angela M. Gurnell
School of Geography
Queen Mary University of London
London E1 4NS, UK
a.m.gurnell@qmul.ac.uk

Marwan A. Hassan
Department of Geography
University of British Columbia
Vancouver, BC, Canada V6T 1Z2
mhassan@geog.ubc.ca

D. Murray Hicks
NIWA
Christchurch, New Zealand
m.hicks@niwa.cri.nz

Cliff R. Hupp
US Geological Survey
Reston, VA 20192, USA
crhupp@usgs.gov

Robert B. Jacobson
US Geological Survey
Columbia, MO 65201, USA
rjacobson@usgs.gov

L. Allen James
Department of Geography
University of South Carolina
Columbia, SC 29208, USA
ajames@sc.edu

Ichiro Kimura
Department of Civil and Environmental Engineering
Hokkaido University
Sapporo, Japan
i-kimu2@eng.hokudai.ac.jp

G. Mathias Kondolf
Department of Landscape Architecture and Environmental Planning
University of California, Berkeley
Berkeley, CA 94720, USA
kondolf@berkeley.edu

Jessica L. Kozarek
St Anthony Falls Laboratory
University of Minnesota
Minneapolis, MN 55414, USA
jkozarek@umn.edu

Thomas E. Lisle
2512 Cochran Road
McKinleyville, CA 95519, USA
thomas.lisle@gmail.com

Richard R. McDonald
Geomorphology and Sediment Transport Laboratory
US Geological Survey
Golden, CO 80403, USA
rmcd@usgs.gov

François Métivier
Institut de Physique du Globe de Paris
Université Paris 7 – Denis Diderot
Paris, France
metivier@ipgp.fr

David R. Montgomery
Department of Geological Sciences
University of Washington
Seattle, WA 98195, USA
dave@geology.washington.edu

Mohamed Nabi
Department of Civil and Environmental Engineering
Hokkaido University
Sapporo, Japan
m.nabi@eng.hokudai.ac.jp

Jonathan M. Nelson
Geomorphology and Sediment Transport Laboratory
US Geological Survey
Golden, CO 80403, USA
jmn@usgs.gov

James E. O'Connor
US Geological Survey
Portland, OR 97201, USA
oconnor@usgs.gov

Takashi Oguchi
Center for Spatial Information Science
University of Tokyo
Kashiwa 277-8568, Japan
oguchi@csis.u-tokyo.ac.jp

Chris Paola
St Anthony Falls Laboratory
University of Minnesota
Minneapolis, MN 55414, USA
cpaola@umn.edu

François Petit
Institut de Géographie
Université de Liège
4000 Liège, Belgium
francois.petit@ulg.ac.be

Hervé Piégay
EVS - UMR 5600 CNRS
Site ENS Lyon, Université de Lyon
69362 Lyon, France
herve.piegay@ens-lyon.fr

James E. Pizzuto
Department of Geological Sciences
University of Delaware
Newark, DE 19716, USA
pizzuto@udel.edu

Rafael Real de Asua
Stillwater Sciences
Berkeley, CA 94705, USA
raf@stillwatersci.com

Leslie M. Reid
Redwood Sciences Laboratory
USDA Forest Service Pacific Southwest Research Station
1700 Bayview Drive
Arcata, CA 95521, USA
lreid@fs.fed.us

Massimo Rinaldi
Dipartimento di Scienze della Terra
Università degli Studi di Firenze
50139 Florence, Italy
mrinaldi@dicea.unifi.it

Yannick Y. Rousseau
Department of Geography
University of Western Ontario
London, ON, Canada N6A 5C2
yroussea@uwo.ca

André G. Roy
Department of Geography, Planning and Environment
Concordia University
Montréal, QC, Canada H3G 1M8
andreg.roy@concordia.ca

Laurent Schmitt
Laboratoire Image, Ville, Environnement UMR 7362 Unistra–CNRS–ENGEES
Université de Strasbourg
67083 Strasbourg, France
laurent.schmitt@unistra.fr

David A. Sear
Department of Geography
University of Southampton
Southampton SO17 1BJ, UK
d.sear@soton.ac.uk

Yasuyuki Shimizu
Department of Civil and Environmental Engineering
Hokkaido University
Sapporo, Japan
yasu@eng.hokudai.ac.jp

Andrew Simon
Cardno ENTRIX
Oxford, MS 38655, USA
andrew.simon@cardno.com

Michal Tal
CEREGE UMR 7330, 80
Université Aix-Marseille
13545 Aix-en-Provence, France
tal@cerege.fr

Marco J. Van de Wiel
Centre for Agroecology, Water and Resilience
Coventry University
Coventry CV1 5FB, UK
marco.vandewiel@coventry.ac.uk

Lise Vaudor
EVS - UMR 5600 CNRS
Site ENS de Lyon, Université de Lyon
69362 Lyon, France
lise.vaudor@ens-lyon.fr

Des E. Walling
Geography, College of Life and Environmental Sciences
University of Exeter
Rennes Drive
Exeter EX4 4RJ, UK
d.e.walling@exeter.ac.uk

Peter J. Whiting
Department of Earth, Environmental and Planetary Sciences
Case Western Reserve University
Cleveland, OH 44106, USA
peter.whiting@cwru.edu

Series Foreword

Advancing River Restoration and Management

The field of river restoration and management has evolved enormously in recent decades, driven largely by increased recognition of the ecological values, river functions and ecosystem services. Many conventional river management techniques, emphasizing strong structural controls, have proven difficult to maintain over time, resulting in sometimes spectacular failures, and often a degraded river environment. More sustainable results are likely from a holistic framework, which requires viewing the 'problem' at a larger catchment scale and involves the application of tools from diverse fields. Success often hinges on understanding the sometimes complex interactions among physical, ecological and social processes.

Thus, effective river restoration and management require nurturing the interdisciplinary conversation, testing and refining of our scientific theories, reducing uncertainties, designing future scenarios for evaluating the best options, and better understanding the divide between nature and culture that conditions human actions. It also implies that scientists should communicate better with managers and practitioners, so that new insights from research can guide management, and so that results from implemented projects can, in turn, inform research directions.

This series provides a forum for 'integrative sciences' to improve rivers. It highlights innovative approaches, from the underlying science, concepts, methodologies, new technologies and new practices, to help managers and scientists alike improve our understanding of river processes, and to inform our efforts to steward and restore our fluvial resources better for a more harmonious coexistence of humans with their fluvial environment.

G. Mathias Kondolf,
University of California, Berkeley

Hervé Piégay
University of Lyon, CNRS

Preface to the Second Edition

Since the publication of the first edition of *Tools in Fluvial Geomorphology* in 2003, the field has been in the course of a revolution sparked by the development of new tools such as improved remote sensing data, acoustic Doppler profilers and radiometric dating methods. The field has arguably entered a new era in knowledge production, the emergence of a second period of active quantification, likely to have similarly profound impacts as the quantitative revolution of the 1960s. While traditional cross-section surveys and bed material sampling still have their place, analysis of drone-based photogrammetry and GIS analysis of large data sets can yield insights that allow the researcher to see the 'forest' beyond the individual 'trees' knowable from field work at the reach scale.

Moreover, the role of fluvial geomorphology within society is changing, as geomorphologists are increasingly called upon to provide input into ecological assessments, sustainable management and restoration schemes. Sometimes, geomorphology is applied by non-geomorphologists, summarized to simple rules of thumbs, misused, and results misinterpreted. The discipline is fairly rich in terms of techniques available and conceptual background. Practitioners can benefit from a broader array of tools if they understand the full range of methods available and the context of their use in an integrative perspective.

By virtue of its position at the intersection of geography, geology, hydrology, river engineering and ecology, fluvial geomorphology is an inherently interdisciplinary field. The tools used reflect this diversity of backgrounds, with techniques borrowed from these different fields. This diversity is now compounded by the new tools available thanks to recent technological innovations, and by the new demands placed on the field. Thus, the need to update *Tools* to provide a reference work for scientists in allied fields, managers seeking guidance on what kind of geomorphic study is best suited to their needs and students seeking to make sense of the plethora of approaches coexisting within fluvial geomorphology. Geomorphic studies based on this large set of knowledge, and placed within an integrative and interdisciplinary perspective, are more likely to solve the often complex problems faced today.

Most of us are familiar and comfortable with a fairly narrow range of tools. Even if we are not 'one-trick ponies', if left to our own devices, we are still likely to fall back on a small set of more familiar methods of study. The problem is summed up in the popular expression, 'If your only tool is a hammer, every problem looks like a nail'. To enlarge our toolboxes, it can be helpful to have a reference that succinctly summarizes the techniques of specializations other than our own, to help understand the kinds of problems to which different methods are best adapted, and the advantages and disadvantages of each. That is the goal of this book. As we were frequently reminded by the late Reds Wolman, who contributed to the first edition and who provided much of the inspiration for both editions, '*Let the punishment fit the crime*'. That is, use a tool that is well adapted to the specific problem. This requires some understanding of the range of tools available to us, which this book attempts to convey.

We are indebted to our contributors, acknowledged experts in their specific fields, all of whom endeavoured to explain in plain English the workings and pros and cons of various methods in their fields. We thank them for their thoughtful contributions and hope that the book as a whole will encourage readers to expand horizons and integrate geomorphologists' knowledge and know-how in their practices.

Matt Kondolf
Hervé Piégay

SECTION I

Background

CHAPTER 1

Tools in fluvial geomorphology: problem statement and recent practice

G. Mathias Kondolf[1] and Hervé Piégay[2]
[1]University of California, Berkeley, CA, USA
[2]Université de Lyon, UMR 5600 CNRS, Lyon, France

Let the punishment fit the crime.

Popular saying invoked by the late M.G. Wolman during drafting of the first edition of *Tools in Fluvial Geomorphology* to capture the idea that the tools should be selected based on the problem to be solved.

1.1 Introduction

As noted by Wolman (1995), in his essay *Play: the handmaiden of work*, much geomorphological research is applied. The spatial and temporal scales of geomorphic analysis can provide insights for the management of risk from natural hazards, solving problems in river engineering (Giardino and Marston 1999) and river ecology (Brookes and Shields 1996), with recent developments in river restoration in terms of assessment, design and monitoring (Morandi *et al.* 2014). As do all scientists, fluvial geomorphologists employ tools in their research, but the range of tools is probably broader in this field than others because of its position at the intersection of geology, geography and river engineering, which draws upon fields such as hydrology, chemistry, physics, ecology and human and natural history. Increasingly, the tools of fluvial geomorphology have been adopted, used and sometimes modified by non-geomorphologists, such as scientists in allied fields seeking to incorporate geomorphic approaches in their work, managers who prescribe a specific tool be used in a given study, and consultants seeking to package geomorphology in an easy-to-swallow capsule for their clients.

Frequently, a lack of geomorphic perspective shows in the questions posed, which are often at spatial and temporal scales smaller than the underlying cause of the problem. For example, to address complaints about bank erosion problem, we have frequently seen costly structures built to alter flow patterns within the channel. Although the designers may have employed hydraulic formulae to design the structures, they may have neglected to look at geomorphic processes at the basin scale, even at reach scale, so that the driving factors are not well identified. Intervening on the symptoms rather than on the underlying disease itself is usually not the best option to solve problems. In such a case, controlling bank erosion through mechanical means will at best provide only temporary and local relief from a system-wide trend. Moreover, it is now well understood that bank erosion and deposition are essential processes to create the complex and diverse channel (Florsheim *et al.* 2008) and floodplain (Stanford *et al.* 2005) habitats needed by many valued species. Thus, what is seen locally as a problem by a riparian landowner may simply be part of the naturally dynamic river behaviour that supports river ecology, and if bank erosion has increased due to catchment-wide changes, even applying geomorphic tools at the site scale only will ultimately prove ineffective (or at least not sustainable) and ecologically detrimental, because the question was poorly posed at the outset without any robust diagnosis and geomorphic expertise based on the range of available tools.

The purpose of this book is to review the range of tools employed by geomorphologists and to link clearly the choice of tools to the question posed, thereby providing guidance to scientists in allied fields and to practitioners about the sorts of methods available to address questions in the field and the relative advantages and disadvantages of each. This book is the result of a collective effort, involving contributors with diverse ages, disciplinary expertise, professional experience and geographic origins to illustrate the range of tools in the field and their application to problems in other fields or management problems. This second edition has incorporated substantial updates, involving new authors with significant contributions to the field over the past decade.

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

1.2 Tools and fluvial geomorphology: the terms

Webster's Dictionary defines a tool as anything used for accomplishing a task or purpose (Random House 1996). By a tool, we refer comprehensively to concepts, theories, methods and techniques. The distinction among these terms is not always clear, depending on the level of thinking and abstraction. Moreover, definitions vary somewhat with dictionaries (e.g. Merriam 1959 versus Random House 1996) and definitions of one term may include the other terms. In our usage, a concept is defined as a mental representation of a reality and a theory is an explicit formulation of relationships among concepts. Both are tools because they provide the framework within which problems are approached and techniques and methods deployed. A method involves an approach, a set of steps taken to solve a problem and would often include more than one technique. As suggested by *Webster's Dictionary* (Random House 1996), it is an orderly procedure, or process, regular way or manner of doing something. Techniques are the most concrete and specific tools, referring to discrete actions that yield measurements, observations or analyses.

As an illustration, a researcher can base his approach on the fluvial system theory and, within this general framework, one of the field's seminal concepts, the notion of bankfull discharge as being the dominant/geomorphic discharge. To test the relation between bankfull discharge and dominant discharge, he can proceed step by step, identifying a general methodological protocol, first to determine what is the bankfull discharge, then its frequency. He may survey channel slope and cross-sectional geometry and measure water flow and velocity, or, if field measurements of flow were not possible, he might estimate flow characteristics from the surveyed geometry and hydraulic equations. In the general case, measuring flow in the field can be undertaken using several methods, such as applying a portable weir, salt dilution or current meter method, but the former are normally better suited for lower flows than the bankfull discharge being studied. The current meter method could be based on various techniques, such as those to measure flow depth and velocity (e.g. using Pryce AA or other current meters, wading with top-setting wading rods or suspending the meter from a cableway or bridge), mechanically improving the cross-section for measurement, accounting for flow angles and sources of turbulence when placing the current meter in the water and estimating the precision of the measurement. Also, given that channel capacity should be related to the long-term flow frequency (Wharton *et al.* 1989), the researcher would normally analyse long-term gauging data (if available for the river being studied), or synthesize from nearby gauges in the region.

Whereas some tools are specific to fluvial geomorphology, others are borrowed from sister disciplines and some (such as mathematical modelling, statistical analysis and inductive or hypothetico-deductive reasoning) are used by virtually all sciences (Bauer 1996; Osterkamp and Hupp 1996). Compared with many other disciplines, fluvial geomorphology has had a strong basis in field observation and measurement. Even with increased reliance on remote sensing and laboratory analysis, the field component is likely to remain critically important to fluvial geomorphology. In this book, our aim is not to describe generic tools, but to focus on tools currently used by fluvial geomorphologists.

We define fluvial geomorphology in its broadest sense, considering channel forms and processes and interactions among channel, floodplain, network and catchment. A catchment-scale perspective, at least at a network level, is needed to understand channel form and adjustments over time. Of particular relevance are links among various components of the fluvial system, controlling the transfer of water and sediment, states of equilibrium or disequilibrium, reflecting changes in climate, tectonic activity and human effects, over time-scales from Pleistocene (or earlier) to the present. Accordingly, to understand rivers can involve multiple questions and require the application of multiple methods and data sources. As a consequence, we consider fluvial geomorphology at different spatial and temporal scales, within a nested systems perspective (Schumm 1977). Analysis of fluvial geomorphology can involve the application of various approaches from reductionism to a holistic perspective, two extremes of a continuum of underlying scientific approach along which the scientist can choose tools according to the question posed.

1.3 What is a tool in fluvial geomorphology?

Roots and tools

Fluvial geomorphology being at the frontier of several disciplines, the choice of tools is fairly large and benefits from the multiple influences of the training of the investigators. The geologically trained fluvial geomorphologist may be more likely to apply tools such as new techniques of dating such as OSL (optical stimulated luminescence) or isotopes (U/Th isotopic ratios, ^{14}C, ^{137}Cs and ^{210}Pb) and techniques that provide subsurface information (e.g. ground-penetrating radar). By contrast, the investigator trained in river hydraulics and physics is more likely to apply tools such as numerical modelling, flume experiment and mechanics. Some geographers focus on spatial complexity, interactions of fluvial forms and processes according to the characters of the basin or bioclimatic regions within which they are observed, the influence of human activities, vegetation cover, or geological settings, employing tools such as remote sensing, GIS or statistics and field metrology.

Within fluvial geomorphology, different branches are also observed, with researchers tending to focus either on a historical perspective (palaeoenvironmental studies) or on processes (dynamic or functional geomorphology). Interactions with biology are reflected in the term *biogeomorphology* (Viles 1988; Gregory 1992) or *ecogeomorphology* (Frothingham *et al.* 2002;

Thoms and Parsons 2002) for this branch of the discipline. Different journals illustrate this diversity of perspectives, each with a specific focus, such as *Geological Society of America Bulletin*, *Water Resources Research*, *Earth Surface Processes and Landforms*, and *Geomorphology*.

Holistic investigation of rivers requires a multidisciplinary approach. Thus, fluvial geomorphology increasingly interacts with other disciplines such as engineering (e.g. Thorne *et al.* 1997; Gilvear 1999), ecology (Hupp *et al.* 1995), environmental science and management (e.g. Brookes 1995; Thorne and Thompson 1995) and societal issues (Kondolf and Piegay 2011), and is recognized as a key element in river restoration (Wohl *et al.* 2005; Simon *et al.* 2011). These interactions are two-way, in that not only is geomorphology applied to these allied fields, but also tools from the allied fields are applied to fluvial geomorphic problems. Geomorphological techniques, such as grain size sampling and channel facies/habitat assessment, are applied to ecological problems such as assessments of fish habitat, and biological techniques (such as dendrochronology, biochemistry analysis or biometrics) are applied to geomorphological problems, such as dating deposits and surfaces or highlighting variability in forms and processes. More sophisticated statistical analyses developed for understanding complex social or biological objects, are now applied to geomorphic data sets. Likewise, geomorphology's interactions with archaeology have yielded benefits to both fields. As a result of multiple roots and extensive interactions fluvial geomorphology has with other disciplines, the set of tools used in this domain is unusually rich and diverse and many tools are now no longer confined to a single discipline. Useful tools increasingly include airborne and terrestrial LiDAR, satellite and airborne imagery (hyperspectral, hyperspatial and radar) and ground sensors such as radiofrequency identification (RFID) and cameras (Thorndycraft *et al.* 2008; Carbonneau and Piégay 2012).

From conceptual to working tools

As any other discipline, geomorphology is characterized by internal debates about theories and methods used and about its history and development (Smith 1993; Rhoads and Thorn 1996; Yatsu 2002). Amongst the most influential theories have been the cycle of erosion (Davis 1899), concepts of magnitude–frequency and effective discharge (Wolman and Miller 1960) and, more recently, the systems theory which emerged with the quantitative revolution in the 1960s (Church 2010) following the heritage of Gilbert (1877). However, as underlined by Knighton (1984), the field of fluvial geomorphology has developed relatively few original theories, tending rather to import theories from allied fields, such as hierarchical theory, system theory, chaos theory and their associated concepts.

Among methods used in this interdisciplinary field, we can distinguish methods of thought that structure the way we do research and working methods used during the research process, each with its specific techniques (Table 1.1). The inductive method involves generalizations developed from a set of observations. For example, in historical geomorphology, we do not know in advance what we will find, so the field data (e.g. date of deposits provided by archaeological artifacts) drive the research. As another example, the empirical relationships established between the fluvial forms and flow regime have led to the formulation of many new scientific questions. As empirical data have accumulated, the conceptual models of flow-channel form relations have been modified based on the new findings. In contrast, in the deductive method, the research process is driven by a preliminary hypothesis, which may be invalidated, using experimentations and usually statistical tests. The deductive approach can be purely experimental, with the researcher reducing artificially, in laboratory or field, the number of acting variables, to establish and validate the basic links among some of them. It can be based also on comparisons between

Table 1.1 A few examples of thought and working methods.

Thought methods

Inductive method. Generalization from data collected in the field, laboratory, literature, etc. Often an exploratory method, from which some hypotheses can be developed.

Hypothetico-deductive method. A preliminary hypothesis or conceptual model is modified, confirmed or rejected based on results of the (usually field or laboratory) studies. It can be applied by using comparative methods or experimental ones:

- *Systemic/comparative methods.* Simultaneous observations of multiple rivers or reaches, sometimes at multiple scales and involving different levels of a drainage network, from which the scientist attempts to identify distinct forms or types of functioning, sensitivity to changes and potential thresholds. A pair of spatial objects (one control and one observed) is the basic step of doing natural experiments. Working methods and associated techniques developed in inductive approaches can be used, but a preliminary hypothesis has been posed and is tested.
- *Experimental methods.* Controlled conditions are created in the laboratory (e.g. flume) or in the field when possible (e.g. erosion plots with artificial rain). This approach is based on a specific framework of working methods and associated techniques.

Working methods and associated techniques

Pre-field methods. Any approach developed in a preliminary step to select the thought methods and design a data collection framework, in some cases a sampling protocol. Based on the question posed, when, where, and what one does in the field.

Field methods. Any approach to measure processes, forms, and deposits in the field or to collect archival data or any spatial information.

Laboratory methods. Any approach performed in a laboratory on field samples to measure physical, chemical and biotic characteristics.

Post-field methods. Any approach used to treat the data and interpret the results.

spatial objects whose existing conditions are used for testing and validating an *a priori* hypothesis (*in natura* experience) for which both specific areas and specific data are selected.

A restrictive definition of science, such as proposed by Claude Bernard (1890), which excludes humanities and requires a strict trinome of hypothesis, experience and conclusion applied to a simple or simplified object, does not apply well to geomorphology. Laboratory experiments are often used in fluvial geomorphology to complement field studies, but controlled experimentation in the manner of pure physics is not possible for most geomorphological concerns (Baker 1996). More fundamentally, some geomorphic questions cannot be solved by testing of hypotheses posed *a priori* and complex new questions have emerged that we cannot simplify without losing relevance. Similarly, problems are brought to geomorphologists from other fields, problems that are frequently posed at spatial and temporal scales smaller and shorter than those needed to understand the fluvial processes involved.

By virtue of their complexity, fluvial systems can be explored by a set of approaches. With the development of new technologies and larger databases, it is now possible to pose new questions at different spatial levels. It becomes possible to consider complexity and to work with convergence of evidence instead of conclusive proofs, comparisons among multiple sites instead of between treated and control sites, and enlargement of the idea of experimentation to include directed, organized observations over large numbers of sites, partial models (accepting that it is impossible to model the whole system fully) and clearly articulated conceptual models. Comparative analysis becomes increasingly important, especially to consider geomorphological questions holistically.

In this context, there is a clear challenge to mix holistic and reductionist approaches, the first to integrate the studied object in its temporal and spatial context, the second to highlight the physical laws controlling the forms and processes. The inductive and deductive methods can be complementary and, by using both, one can avoid problems of overgeneralizing on the one hand, or reaching conclusions that are only narrowly applicable on the other. Experimentation, conducted in tandem with field observation, can significantly advance our understanding of process (Schumm *et al.* 1987). Over the last decade, a new quantitative revolution occurred with the emergence of new sensors, imaging techniques and computer facilities (Thorndycraft *et al.* 2008; Piégay *et al.* 2015), with the emergence of what some consider to be new sub-disciplines, such as remote sensing of rivers (Marcus and Fonstad 2010) made possible by technological advances in optics, mechanics, electronics, geoinformatics, geocomputing, geopositioning and statistics (Anbazhagan *et al.* 2011). In widening the space and time framework, these new analyses and resultant data sets improve our understanding of how local observations can be generalized, how channel states are variables in time and more closely connect reductionist and holistic approaches to understanding the complexity of geomorphic processes.

Multidisciplinary approaches, such as coupling hydraulics and geomorphology, have facilitated the application of physics and mechanics to the field. This has resulted in better understanding of the acting processes, limits of validity of given laws and limitations of numerical models. Using bank erosion as an example, geomorphological research has identified the complexity of geographical contexts and physical processes controlling bank erosion, its important ecological role, potential consequences of hard bank protection and alternative solutions to perceived erosion problems, such as erodible corridors, the implementation of which requires an interdisciplinary approach, e.g. with legal scholars to address property rights, sociologists to collect opinions of landowners and economists to evaluate the long-term economics of various alternatives. This evolution of the research perspective has been accompanied by increasing participation in decision-making by citizens, landowners, governmental and non-governmental agencies and other stakeholders.

Working methods are diverse, because there are many ways of approaching fluvial geomorphological questions, in the field, remotely sensed from airborne and satellite or experimentally, from archives and historical data, at various spatial and temporal scales, in various man-made and natural contexts. We propose a rough classification based on the stage at which the methods are used: pre-field, field and post-field methods, with 'field' being considered here in a larger sense not only of data collection in the landscape but also in archives, airborne/satellite surveys, and so on (Table 1.1). Sampling methods, sites, frequency, and so on, must be frequently determined before collecting data. Once these preliminary questions have been answered, methods are developed to extract information from existing data and images or collect information in the field, potentially reinforced by laboratory techniques to measure quantities, concentrations or dates. At the post-field stage, other methods (e.g. statistical, graphics, mapping, imagery analysis) are used to treat the data and interpret the results. Whatever the stage, the methods and techniques used depend strongly on the question posed and the thought method chosen, whether to describe, to explain, to simulate or to predict.

The organization in Table 1.1 is obviously only one of many ways to classify these, but it provides an overview of current approaches in the discipline. Under each working method (as defined in Table 1.1.), a number of techniques may be used, depending on the characteristics of the field site and the nature of the question posed. For example, there are multiple methods for measuring discharge, one of which is the current meter method, involving measurements of depth and velocity across the channel, and another being the salt dilution method (Chapter 12). The method of bedload sampling can involve techniques such as bedload traps, Helley–Smith sampling or tracer gravels (Chapter 13). However, the line between method and technique is not always clear, as the more one knows about a tool and its components and variants, the more one is inclined to call it a method rather than a technique. For example, to the non-specialist, dating or assessing overbank sedimentation rates from ^{137}Cs concentration measurement in the soil profile of a floodplain appears at first to be a technique, but to the specialist it is a method, which can involve several techniques, such as sampling (from coring and slice-cutting to obtain sediment samples, digging bulk samples or profile analysis), as well as measuring radioactivity (high- versus low-resolution spectrometer, alpha versus gamma spectrometry) (Chapter 9).

In this book, we focus not only on the field/laboratory methods and techniques as they refer to the specific tools of the discipline but also to key concepts and methods that are fundamental for the geomorphological thinking, the way of approaching the applied problems. Because pre- and post-field methods and techniques are more generic tools in science, we focus less on these. Moreover, we have organized the book according to key geomorphological topics rather than to key tools because one of our main messages is that the geomorphological question is key. The tools themselves are secondary and follow directly from the question. Accordingly, we introduce the tools by the question posed, considering five main types of geomorphological questions and then associated tools:

- *the historical framework* and the methods and associated techniques to date and assess historical geomorphological trends;
- *the spatial framework* and the concepts, methods and associated techniques that reveal spatial structure and nested character of fluvial forms;
- *the chemical, physical and biological methods* for dating and the study of spatial structure and fluvial processes;
- *the analysis of processes and forms*, the traditional heart of the discipline based on field surveys and measurements of sediment and water flow;
- *the future framework* for which methods and techniques exist for discriminating, simulating and modelling processes and trends.

The aim is not to describe specific techniques in detail, but rather to focus on the geomorphological methods within which techniques are applied. Techniques have been well described in specific papers, and also in more comprehensive works (e.g. Dackcombe and Gardiner 1983; Thorne 1998; Goudie *et al.* 2005; Sear *et al.* 2010). The greatest contribution of this book, then, is probably to develop better the context within which the different tools are chosen and to enrich the description of methods and techniques by contrasted examples. Two chapters are also specifically devoted to conceptual approaches, such as the fluvial system theory and the sediment budget concept. Through these treatments, we seek to show the manner and spirit in which the geomorphologist works.

Tools and questions

Concepts, theories, methods and techniques are tools used to answer questions (Fig. 1.1). The key element, then, is the question. This is true even for an inductive approach, because there is an implicit question posed of what kind of geomorphological forms and processes trends occur. The efficacy of geomorphic research depends much less on the choice of method than on the quality of the research question posed (Leopold and Langbein 1963). Once the question is posed, based on deductive or inductive approaches and supported by a given set of concepts, supposing that it is valid, the second step is to define the working methods and potential data sources. Next (or simultaneously), methods and associated techniques are identified within a given conceptual framework and with spatial and temporal resolution appropriate to the scale at which the question is posed. When one considers the river as a system, one's questions are usually less time and scale restrictive and one tends to pose specific questions about links among catchment sub-divisions.

Concepts can be both the result of a research programme (question → result → generalization → concepts and theories) and also tools with which to carry out the research, as once established, concepts allow us to organize data and guide our subsequent research. The graded river concept (Mackin 1948), the concept of dynamic equilibrium (Hack 1960; Chorley and Kennedy 1971) and the concept of reaction and relaxation times (Graf 1977; Brunsden 1980) are all a result of generalization provided by previous research. These concepts also led to the development of other research questions, which in turn were tested in various environments in order to understand better the sensitivity of regional contexts and the variability of thresholds. However, as with other tools, concepts may be applicable in some situations but not in others. For example, the concept of dominant discharge as a frequently occurring flood (such as the $Q_{1.5}$) is a useful concept in humid climates or snowmelt streams, but generally is of little use in semi-arid climate channels (Wolman and Gerson 1978) or braided alpine rivers (Belletti *et al.* 2014). We can also step back to larger scale conceptual frameworks or conceptual models that guide our research, the continuum concept (Leopold and Wolman 1957; Vannote *et al.* 1980), the fluvial system concept (Schumm 1977), the hydrosystem concept (Roux 1982) and the sediment budget (Dietrich and Dunne 1978).

In most cases, there is no perfect tool to answer the question posed. Instead, we usually must employ a set of (often diverse) them to approach a question. Ironically, it seems that non-geomorphologists sometimes assume that a perfect tool must exist to answer their questions easily and thus they may readily adopt tools whose proponents claim are ideally suited to address management concerns. For example, Bevenger and King (1997, p. 1393) argued that there was a need for 'well designed monitoring protocols' using 'tools that are relatively simple to implement, that can be used directly and consistently by field personnel and are sensitive enough to provide a measure of impact'. While probably no-one would disagree with the desirability of such protocols, there is no *a priori* reason to assume that they must exist and certainly the 'zig-zag' method of sampling bed material promoted by Bevenger and King

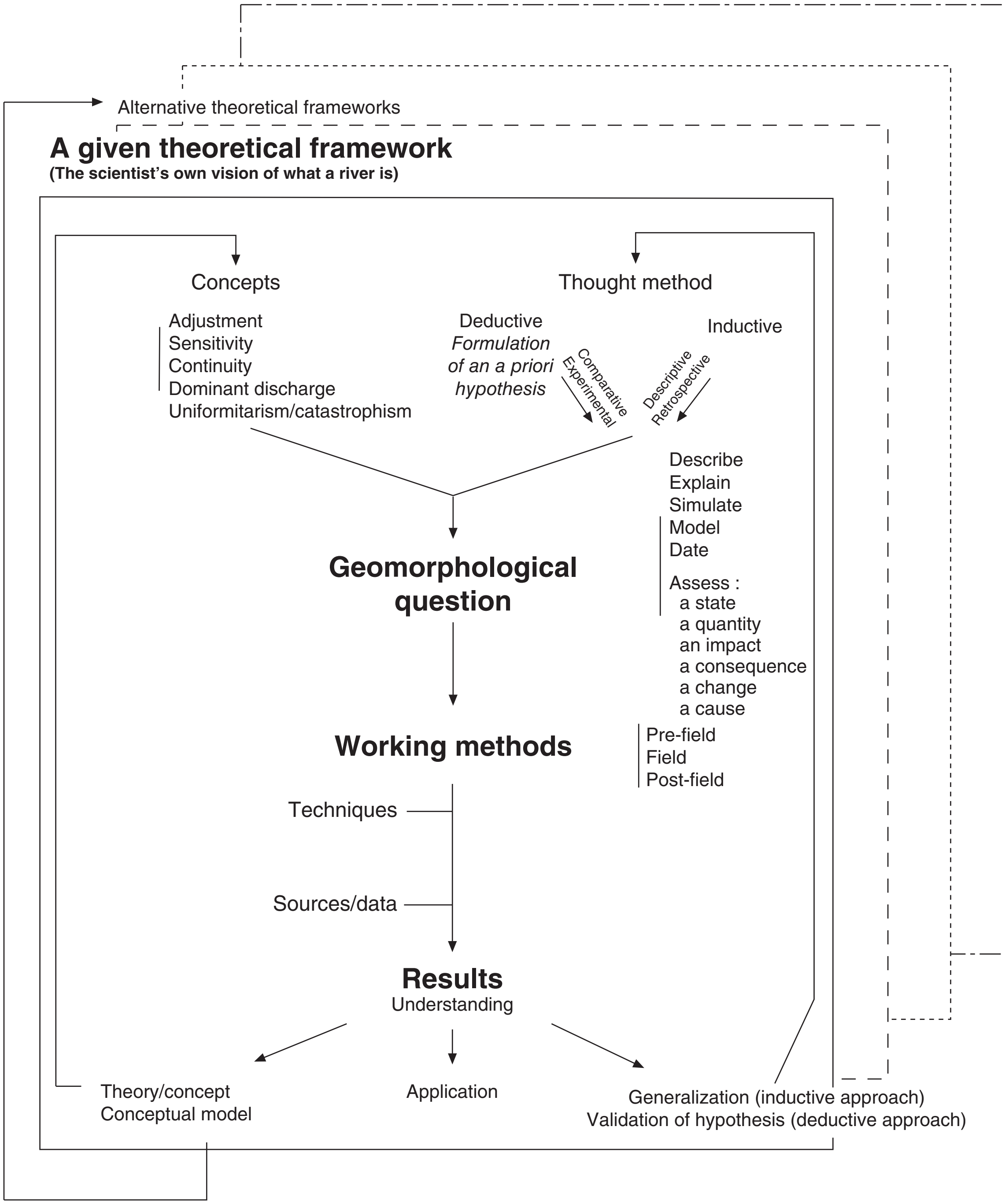


Figure 1.1 General framework of the way in which a geomorphological question is posed in the research process, and use of different tools: theoretical framework, concepts, thought methods and working methods, with their associated techniques variously dependent on the sources used.

(1995, 1997) fell far short of such an ideal (Kondolf 1997a,b; see discussion in Chapter 13 of this volume).

We posit a relation among theories, data sources from field, laboratory or satellite and airborne sensors and various kinds of knowledge that highlights process understanding from field observations and measures combined with numerical modelling or temporal change analysis, all depending on advances in data production and methodology partly related to external inputs. We can also recognize motivations for knowledge production versus practical delivery needed to inform activities such as river restoration (Fig. 1.2.).

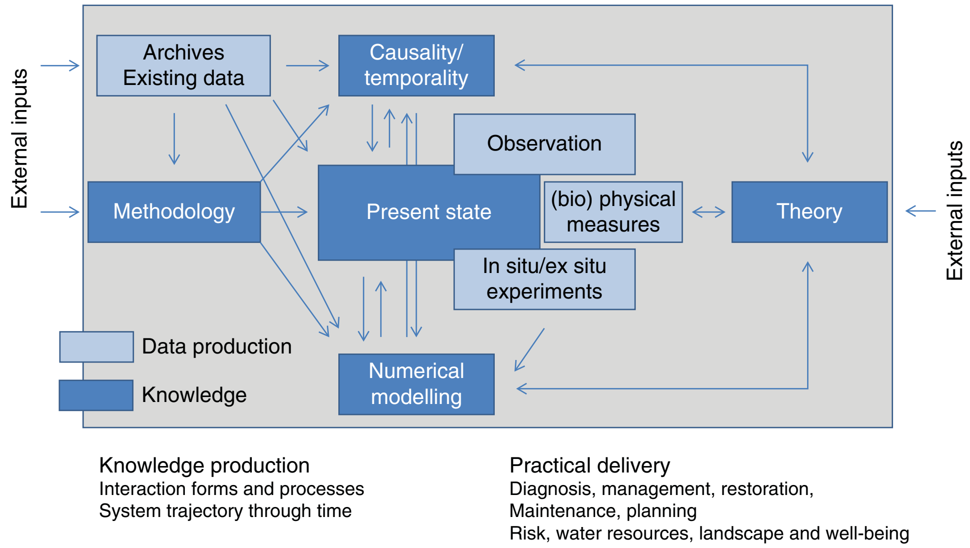


Figure 1.2 Summary of the different approaches conducted to answer geomorphic problems based on data and knowledge production.

1.4 Overview and trends of tools used in the field

In the first edition of *Tools in Fluvial Geomorphology*, we presented a quantitative analysis of papers published in the field from 1987 to 1997 in the journals *Catena*, *Earth Surface Processes and Landforms* (*ESPL*), *Géographie Physique et Quaternaire* (*GpQ*), *Geomorphology* and *Zeitschrift für Geomorphologie* (*ZfG*). More recently, Piégay *et al.* (2015) extended the analysis forward to capture the years in which new technologies have been extensively adopted and to cover more than two decades (1987–2009). This study documented an increase in publications in fluvial geomorphology, similarly to the increase in geomorphology publications generally, so that fluvial geomorphology usually represent ~30% of the production of the discipline. Authorship has broadened, with increased diversity of the country origin of the first author (as measured by the Shannon index), but almost half of the authors are from anglophone countries: the United States and United Kingdom, with 28 and 20% of contributions, respectively, followed by Canada (9%) and Australia and New Zealand (7.2%). Countries such as France and China significantly increased in importance after 2000.

The most popular spatial scale was the channel (with 62% of papers), followed by the basin/network (18%) and floodplain (11%). When considering the time-scale covered, 51% of papers deal with present time, 16% with decadal and others (no time-scale, century, holocene, quaternary) dealing with ~6–8% each. The frequency of contributions on present was stable between 1986 and 2009 whereas papers dealing with decadal scale increased after 2000 (Piégay *et al.* 2015).

The increase in publication and diversity of participation in the field has come at a time of tremendous change in technology (as noted above), further diversifying the choice of tools used by workers in fluvial geomorphology. Looking at the tools reported in papers published in the journals *ESPL* and *Geomorphology* (which published the majority of fluvial geomorphic papers), field measurements (such as cross-section surveys and other usually reach-scale measures) constituted the most popular 'tool' from a list of 53 different tools identified in table 2 and fig. 9 in Piégay *et al.* (2015). Other commonly employed tools included other field measures such as grain size and associated use of hydraulic formulae and geomorphic mapping, archived and aerial photo analysis, DEM and GIS analysis.

The data show a significant change in methods used in the field as a result of the increase in data availability and new sources of information from remote field sensing (ground, airborne and satellite). Clearly, a new era in knowledge production is observed since 2000, showing the emergence of a second period of active quantification and an internationalization of the fields opening new scientific questions because the diverse scientific traditions of different countries combine to offer fresh perspectives. A new debate is then emerging on the field tradition of geomorphology because the methodological revolution implies fundamental changes to scientific practices and opens new issues in terms of knowledge production.

1.5 Scope and organization of this book

As suggested by the literature review, the multiple disciplinary roots of fluvial geomorphology, the field's increasing interaction with other disciplines and applications to management problems and the availability of new technologies with transformative potential over the past two decades have resulted in an array of tools that is diverse and becoming more so. This book presents summaries of the tools used in various areas of fluvial geomorphology, written at a level that falls between broad generalization and highly specific instruction on technique. The aim of the chapters is to help managers or scientists in other fields (or simply other sub-fields in fluvial geomorphology) understand the capabilities and limitations of various geomorphic tools, to aid in choosing methods appropriate to the questions posed. Of course, this requires an understanding of how various tools fit within the conceptual framework of the field ('big picture' context) and it requires some explanation of how the methods are actually carried out, the equipment and resources required, accuracy and precision, and so on. For detailed instructions and descriptions of the equipment and supplies needed, we refer the reader to more specialized works. Most chapters include case studies to illustrate applications of the tools described. Although the scope of this book is broad, it does not cover geophysical methods, simply to limit the book to a more manageable scope.

This book is organized into an introduction, five topical sections each with two to five chapters and a conclusion. Following the present introductory chapter, the second section concerns the temporal framework, moving from mainly physical evidence and longer time-scales to more recent and anthropic evidence. The section begins with Chapter 2, in which Robb Jacobson, Jim O'Connor and Takashi Oguchi review surficial geological tools, such as floodplain stratigraphy and slackwater deposits, from which past hydrologic and geomorphic events (such as floods) can be interpreted and dated, changes in land use inferred, and so on. In Chapter 3, Tony Brown, François Petit and Allen James discuss the use of archaeology and human artifacts (such as mining waste) to measure and date fluvial geomorphic processes and events. In Chapter 4, Robert Grabowski and Angela Gurnell review the use of historical records to document and date geomorphic change, mostly in recent centuries and decades.

The next (third) section addresses the spatial framework, emphasizing spatial structure and the nested character of fluvial systems. In Chapter 5, Hervé Piégay reviews the systems approach in fluvial geomorphology, from its roots in strictly physical processes through more recent systems approaches that integrate ecological processes. In Chapter 6, David Gilvear and Robert Bryant review the applications of aerial photography and other remotely sensed data to fluvial geomorphology, from traditional stereoscopic air photo interpretation to more recently developed remote-sensing techniques. In Chapter 7, Matt Kondolf, Hervé Piégay, Laurent Schmitt and David Montgomery review the uses and limitations of geomorphic channel classification systems, tools that have become extremely popular recently among non-geomorphologists, especially as applied to management questions. Concluding the spatial framework section, Peter Downs and Rafael Real de Asua review approaches to modelling catchment processes in Chapter 8.

The fourth section covers chemical, physical and biological evidence, i.e. the applications of methods in these allied field to fluvial geomorphic problems. In Chapter 9, Des Walling and Ian Foster review chemical and physical methods, with a substantial section on isotopic methods for dating, with their revolutionary effect on the field. Cliff Hupp, Simon Dufour and Gudrun Bornette detail biological methods, such as dendrochronology and vegetative evidence of past floods, in Chapter 10.

The fifth section includes analyses of processes and forms. In Chapter 11, Andrew Simon, Janine Castro and Massimo Rinaldi describe methods to analyze channel form, emphasizing field survey and measurement techniques. Peter Whiting details methods of flow and velocity measurement in Chapter 12. In Chapter 13, Matt Kondolf and Tom Lisle review methods of bed sediment measurement (surface and subsurface) in light of various possible research objectives. Tracers, such as painted gravels, magnetic rocks and clasts fitted with radio transmitters, are reviewed by Marwan Hassan and André Roy in Chapter 14. Methods of measuring and calculating sediment transport, suspended, bedload and dissolved, are reviewed by Murray Hicks and Basil Gomez in Chapter 15. Sediment budgets are increasingly used as an organizing framework in fluvial geomorphology, especially in studies of impacts of human actions such as dams. In Chapter 16, Leslie Reid and Tom Dunne draw upon their pioneering work in this area to provide guidance on how to approach sediment budget construction under various objectives and field situations.

The sixth section concerns tools for discriminating, simulating and modelling processes and trends. In Chapter 17, Marco Van de Wiel, Yannick Rousseau and Stephen Darby lay out general considerations for models in fluvial geomorphology. Jon Nelson, Richard McDonald, Yasuyuki Shimizu, Ichiro Kimura, Mohamed Nabi and Kazutake Asahi provide a thorough review of the broad topic of hydraulic and sediment transport modelling methods in Chapter 18. Methods for modelling channel changes are described by Jim Pizzuto in Chapter 19. In Chapter 20, François Métivier, Chris Paola, Jessica Kozarek and Michal Tal review the uses of flume experiments in fluvial geomorphology. In Chapter 21, Hervé Piégay and Lise Vaudor review statistical tools in fluvial geomorphology, not only commonly used tools such as regression, but also statistical analyses often applied in allied fields such as ecology but rarely in geomorphology.

Most of the tools described in this book can be used to answer applied questions and, given the increasing demand for geomorphic input to river management, a wider range of tools deserve to be employed in support of management decisions. The concluding chapter by Hervé Piégay, Matt Kondolf and David Sear (Chapter 22) considers the bridge between geomorphology and management and presents illustrations from the United States, United Kingdom and France of fluvial geomorphology used to help river ecologists, planners and managers to answer to their own questions and, in some cases, to redefine their questions on a larger spatial and temporal scale.

Obviously, a survey of tools in this field could be organized in different ways and even within the chosen structure there are a number of tools that could logically have gone in different chapters and chapters that could have gone in different sections.

For example, ^{137}Cs and ^{210}Pb analyses are usually used in a temporal sense (e.g. to assess the variability of sedimentation rate over time), but they can also be used at a catchment scale to distinguish erosional from depositional areas. Aerial photography can be used to support a range of studies, from historical channel evolution to mapping of spatial patterns over large areas at one point in time. Like ecology or medicine, fluvial geomorphology is a synthesis science, analogous to the composite sciences as visualized by Osterkamp and Hupp (1996), meaning that it is based on a range of tools. Fluvial geomorphology is a thematic area where some scientific disciplines can interact and produce real interdisciplinary insights.

As a consequence, we cannot adopt one way of approaching geomorphological problems and neglect all others. In combination, multiple tools can be helpful in appreciating problems and addressing societal needs. Fluvial geomorphology can be useful in river management, especially as managers begin to think at different time and spatial scales (as implied when one adopts sustainability as a goal). Probably all geomorphologists would agree that it is necessary to specify the problem as clearly as possible and to use the most appropriate tools from the great range now available. This book is intended to help in the realization of these aims.

Acknowledgements

Each chapter in this book was peer reviewed by two reviewers, usually one external and one contributor and by the two Editors, with additional reviews commissioned as needed for significantly revised (or new) chapters. We are indebted to the following reviewers who contributed to the improvement of the book: Vic Baker, James Bathurst, Christian Braudrick, Tony Brown, Pierre Clément, John Buffington, Mike Church, Nic Clifford, Peter Downs, Jonathan Friedman, David Gilvear, Ken Gregory, Basil Gomez, Gordon Grant, Angela Gurnell, Jud Harvey, Marwan Hassan, Nicolas Lamouroux, John Laronne, Eric Larsen, Stuart Lane, Fred Liébault, Mike Macklin, Andrew Miller, David Montgomery, Gary Parker, François Petit, Geoffrey Petts, Didier Pont, Michel Pourchet, Ian Reid, Steve Rice, David Sear, Stanley Trimble, Peter Wilcock and Ellen Wohl.

References

Anbazhagan, S., Subramanian, S.K. and Yang, X. 2011. *Geoinformatics in Applied Geomorphology*, Boca Raton, FL: CRC Press.

Baker, V.R. 1996. Hypothesis and geomorphological reasoning. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 57–85.

Bauer, B.O. 1996. Geomorphology, geography and science. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 381–413.

Belletti, B., Dufour, S. and Piégay, H. 2014. Regional assessment of the multi-decadal changes in braided riverscapes following large floods (Example of 12 reaches in South East of France). *Advances in Geosciences* 37: 57–71.

Bernard, C. 1890. *La Science Expérimentale*, 3rd edn., Paris: J.B. Baillière et Fils, 448 pp.

Bevenger, G.S. and King, R.M. 1995. *A Pebble Count Procedure for Assessing Watershed Cumulative Effects, USDA Forest Service Research Paper RM-RP-319*, Fort Collins, CO: Rocky Mountain Forest and Range Experiment Station.

Bevenger, G.S. and King, R.M. 1997. Discussion of "Application of the pebble count: notes on purpose, method, and variants", by G. M. Kondolf. *Journal of the American Water Resources Association* 33(6): 1393–1394.

Brookes, A. 1995. Challenges and objectives for geomorphology in U.K. river management. *Earth Surface Processes and Landforms* 20: 593–610.

Brookes, A. and Shields, F.D. 1996. *River Channel Restoration: Guiding Principles for Sustainable Projects*, Chichester: John Wiley & Sons.

Brunsden, D. 1980. Applicable models of long term landform evolution. *Zeitschrift für Geomorphologie* 36: 16–26.

Carbonneau, P. and Piégay, H. 2012. *Fluvial Remote Sensing for Science and Management*, Chichester: John Wiley & Sons.

Chorley, R.J. and B.A. Kennedy. 1971. *Physical Geography: a Systems Approach*, London: Prentice Hall, 370 pp.

Church, M. 2010. The trajectory of geomorphology. *Progress in Physical Geography* 34: 265–286.

Dackcombe, R.V. and Gardiner, V. 1983. *Geomorphological Field Manual*, London: George Allen and Unwin.

Davis, W.M. 1899. The geographical cycle. *Geographical Journal* 14: 481–504.

Dietrich, W.E. and Dunne, T. 1978. Sediment budget for a small catchment in mountainous terrain. *Zeitschrift für Geomorphologie, Supplement Band* 29: 191–206.

Florsheim, J.L., Mount, J.F. and Chin, A. 2008. Bank erosion as a desirable attribute of rivers. *Bioscience* 58(6): 519–529.

Frothingham, K.M., Rhoads, B.L. and Herricks, E.E. 2002. A multiscale conceptual framework for integrated ecogeomorphological research to support stream naturalization in the agricultural midwest. *Environmental Management* 29: 16–33.

Giardino, J.R. and Marston, R.A., eds. 1999. Changing the Face of the Earth: Engineering Geomorphology. *Geomorphology* 31(1–4): 1–439 (special issue).

Gilbert, G.K. 1877. *Report on the Geology of the Henry Mountains*, Washington, DC: US Geographical and Geological Survey of the Rocky Mountain Region, 160pp.

Gilvear, D.J. 1999. Fluvial geomorphology and river engineering: future roles utilizing a fluvial hydrosystems framework. *Geomorphology* 31: 229–245.

Goudie, A., Anderson, M., Burt, T., Lewin, J., Richards, K., Whalley, B. and Worsley, P. 2005. *Geomorphological Techniques, British Geomorphological Research Group*, 2nd edn., London: Routledge, 592pp.

Graf, W.L. 1977.The rate law in fluvial geomorphology. *American Journal of Science* 277: 178–191.

Gregory, K.J. 1992. Vegetation and river channel processes. In: Boon, P.J., Calow, P. and Petts, G.E., eds., *River Conservation and Management*, Chichester: John Wiley & Sons, pp. 255–269.

Hack, J.T. 1960. Interpretation of erosional topography in humid temperate regions. *American Journal of Science* 258A: 80–97.

Hupp, C.R., Osterkamp, W.R. and Howard, A.D., eds. 1995. *Biogeomorphology – Terrestrial and Freshwater Systems*, Amsterdam: Elsevier Science, 347pp.

Knighton, D. 1984. *Fluvial Forms and Processes*, London: Edward Arnold, 218pp.

Kondolf, G.M. 1997a. Application of the pebble count: reflections on purpose, method, and variants. *Journal of the American Water Resources Association* 33(1): 79–87.

Kondolf, G.M. 1997b. Reply to discussion by Gregory S. Bevenger and Rudy M. King on "Application of the pebble count: reflections on purpose, method, and variants." *Journal of the American Water Resources Association* 33(6): 1395–1396.

Kondolf, G.M. and Piégay, H. 2011. Geomorphology and society. In: Gregory, K.J. and Goudie, A.S., eds., *SAGE Handbook of Geomorphology*, London: SAGE Publications, pp. 105–117.

Leopold, L.B. and Langbein, W.B. 1963. Association and indeterminacy in geomorphology. In: Albritton, C.C., ed., *The Fabric of Geology*, Palo Alto, CA: Cooper and Co., pp. 184–192.

Leopold, L.B. and Wolman, M.G. 1957. River channel patterns; braided, meandering and straight. *U.S. Geological Survey Professional Paper* 282-b: 39–85.

Mackin, J.H. 1948. Concept of the graded river. *Bulletin of the Geological Society of America* 59: 463–512.

Marcus, W.A. and Fonstad, M.A. 2010. Remote sensing of rivers: the emergence of a subdiscipline in the river sciences. *Earth Surface Processes and Landforms* 35: 1867–1872.

Merriam Company, G. & C. 1959. *Webster's New Collegiate Dictionary*, Springfield, MA: G. & C. Merriam.

Morandi, B., Piégay, H., Lamouroux, N. and Vaudor, L. 2014. How is success or failure in river restoration projects evaluated? Feedback from French restoration projects. *Journal of Environmental Management* 137(1): 178–188.

Osterkamp, W.R. and Hupp, C.R. 1996. The evolution of geomorphology, ecology and other composite sciences. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 415–441.

Piégay, H., Kondolf, G.M., Minear, J.T. and Vaudor, L. 2015. Trends in publications in fluvial geomorphology over two decades: a truly new era in the discipline owing to recent technological revolution? *Geomorphology* 248: 489–500.

Random House. 1996. *Webster's Dictionary*, 2nd ed., New York: Ballantine Books.

Rhoads, B.L. and Thorn, C.E. 1996. *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, 481pp.

Roux, A.L., coord. 1982. *Cartographie Polythématique Appliquée à la Gestion Écologique des Eaux: Étude d'un Hydrosystème Fluvial: le Haut-Rhône Français*. Lyon: CNRS-Piren, 113pp.

Schumm, S.A. 1977. *The Fluvial System*. New York: John Wiley & Sons.

Schumm, S.A., Mosley, M.P. and Weaver, W.E. 1987. *Experimental Fluvial Geomorphology*, New York: John Wiley & Sons, 413pp.

Sear, D.A., Newson, M.D. and Thorne, C.R. 2010. *Guidebook of Applied Fluvial Geomorphology*, London: Thomas Telford.

Simon, A., Bennett, S.J. and Castro, J.M., eds. 2011. *Stream Restoration in Dynamic Fluvial Systems: Scientific Approaches, Analyses, and Tools*. Geophysical Monograph Series, vol. 194, Washington, DC: American Geophysical Union, 544pp.

Smith, D.G. 1993. Fluvial geomorphology: where do we go from here? *Geomorphology* 7: 251–262.

Stanford, J.A., Lorang, M.S. and Hauer, F.R. 2005. The shifting habitat mosaic of river ecosystems. *Verhandlungen des Internationalen Verein Limnologie* 29: 123–136.

Thoms, M.C. and Parsons, M. 2002. Eco-geomorphology: an interdisciplinary approach to river science. In: Dyer, F.J., Thoms, M.C. and Olley, J.M., eds., *The Structure, Function and Management Implications of Fluvial Sedimentary Systems (Proceedings of an International Symposium Held at Alice Springs, Australia, September 2002)*, IAHS Publication No. 276, Wallingford: IAHS Press, pp. 113–119.

Thorndycraft, V.R., Benito, G. and Gregory, K.J. 2008. Fluvial geomorphology: A perspective on current status and methods. *Geomorphology* 98: 2–12.

Thorne, C.R. 1998. *Stream Reconnaissance Handbook. Geomorphological Investigation and Analysis of River Channels*, Chichester: John Wiley & Sons, 133pp.

Thorne, C.R. and Thompson, A., eds. 1995. Geomorphology at Work. *Earth Surface Processes and Landforms* 20(7): 583–705 (special issue).

Thorne, C.R., Hey, R.D. and Newson, M.D. 1997. *Applied Fluvial Geomorphology for River Engineering and Management*, Chichester: John Wiley & Sons, 376pp.

Vannote, R.L., Minshall, G.W., Cummins, K.W., Sedell, J.R. and Cushing, C.E. 1980. The river continuum concept. *Canadian Journal of Fisheries and Aquatic Science* 37: 130–137.

Viles, H.A., ed. 1988. *Biogeomorphology*. Oxford: Blackwell.

Wharton, G., Arnell, N.W., Gregory, K.J. and Gurnell, A.M. 1989. River discharge estimated from channel dimensions. *Journal of Hydrology* 106: 365–376.

Wohl, E., Angermeier, P.L., Bledsoe, B., Kondolf, G.M., MacDonnell, L., Merritt, D.M., Palmer, M.A., Poff, N.L. and Tarboton, D. 2005. River restoration. *Water Resources Research* 41: W10301.

Wolman, M.G. 1995. Play: the handmaiden of work. *Earth Surface Processes and Landforms* 20: 585–591.

Wolman, M.G. and Gerson, R. 1978. Relative scales of time and effectiveness of climate in watershed geomorphology. *Earth Surface Processes and Landforms* 3: 189–208.

Wolman, M.G. and Miller, J.P. 1960. Magnitude and frequency of forces in geomorphic processes. *Journal of Geology* 68: 54–74.

Yatsu, E. 2002. *Fantasia in Geomorphology*, Tokyo: Sozosha, 215pp. (reprint of "To make geomorphology more scientific" and its supplemental discussion).

SECTION II

The Temporal Framework: Dating and Assessing Geomorphological Trends

CHAPTER 2

Surficial geological tools in fluvial geomorphology

Robert B. Jacobson[1], Jim E. O'Connor[2] and Takashi Oguchi[3]

[1]US Geological Survey, Columbia, MO, USA
[2]US Geological Survey, Portland, OR, USA
[3]Center for Spatial Information Science, University of Tokyo, Kashiwa, Japan

2.1 Introduction

Environmental scientists are increasingly asked how rivers and streams have been altered by past environmental stresses, whether rivers are subject to physical or chemical hazards, how they can be restored and how they will respond to future environmental changes. These questions present substantive challenges to the discipline of fluvial geomorphology as they require a long-term understanding of river-system dynamics. Complex and non-linear responses of rivers to environmental stresses indicate that synoptic or short-term historical views of rivers will often give an incomplete understanding. Fluvial geomorphologists can address questions involving complex river behaviours by drawing from a tool box that includes the principles and methods of geology applied to the surficial geological record.

A central concept in Earth Sciences holds that 'the present is the key to the past' (Hutton 1788, cited in Chorley *et al.* 1964), that is, understanding of current processes permits the interpretation of past deposits. Similarly, an understanding of the past can be key to predicting the future. A river's depositional history can be indicative of trends or episodic behaviours that can be attributed to particular environmental stresses or forcings. Its history may indicate the role of low-frequency events such as floods or landslides in structuring a river and its floodplain or a river's depositional history can provide an understanding of its natural characteristics to serve as a reference condition for assessments and restoration.

However, the surficial geological record contained in river deposits is incomplete and biased and it presents numerous challenges of interpretation. The stratigraphic record in general has been characterized as '… a lot of holes tied together with sediment' (Ager 1993). Yet this record is critical in the development of integrated understanding of fluvial geomorphology because it provides information that is not available from other sources. The surficial geological record can present information that predates historical observations or is highly complementary to historical records. Although river deposits are rarely complete enough to form precise predictive models, they provide contextual information that can constrain predictions and help guide choices of appropriate processes to study more closely and hypotheses to test (Baker 1996). Floodplain chronicles of Earth history also provide data sets for calibration and verification of predictive geomorphic models.

The purpose of this chapter is to introduce and discuss surficial geological tools that can be used to improve the understanding of fluvial geomorphology. We present general descriptions of geological tools, provide selected field-based examples and discuss the expectations and limitations of geological approaches. We do not attempt to discuss techniques in detail or to review the entire literature on techniques or applications. Instead, our emphasis is on the conceptual basis of how geological tools are used in geomorphological reasoning. The chapter begins with some general descriptions of stratigraphic, sedimentological and pedological tools, followed by examples of how these tools can be applied to geomorphological analysis of fluvial systems.

2.2 Overview of surficial geological approaches

The analysis of deposits left behind by rivers involves approaches from many disciplines, each of which has its own technical lexicon. For clarity, we will begin with some common definitions. *Surficial geology* refers to the study of the rocks and mainly unconsolidated materials that lie at or near the land surface (Ruhe 1975). In our usage, surficial geology includes the application of sedimentology, geochronology, pedology and stratigraphy to studies of surficial deposits and geomorphology. *Alluvium* is the detrital sediment deposited by rivers, ranging from clay size (<0.002 mm) to boulder size (>256 mm) materials, including detrital organic material. Alluvium is used interchangeably with *fluvial deposits*. Alluvium typically occurs on the landscape in modern channel and bar deposits and in deposits that underlie adjacent floodplains, terraces and alluvial fans.

The term *floodplain deposits* will be defined here restrictively to denote deposits adjacent to a river channel that are being

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

deposited under the current hydrological regime, typically by flow events with frequencies of 0.5–1 yr^{-1} and higher (Leopold *et al.* 1964, p. 319). It should be noted, however, that in some environments, floodplains may be primarily constructed by flows of much lower frequency (for example, Baker 1977). *Floodplain* is used by different professions in different ways. Definitions range from the entire valley bottom outside of the channel to specific statistical definitions such as the 100- or 500-year floodplain. In this chapter, *floodplain* refers to the geomorphic surface underlain by floodplain deposits, that is, those sediments being deposited by relatively frequent floods under the current hydrological regime. Other fluvial deposits adjacent to a river will be referred to as *terrace deposits* or, in combination with floodplain deposits, as undifferentiated *valley-bottom alluvium*. The term *terrace* will apply to surfaces of abandoned floodplains (Leopold *et al.* 1964). The term *soil* is used in this chapter in the pedogenic sense: mineral and organic material at the Earth's surface that has been altered by weathering processes and living organisms (Holliday 1990). Hence a floodplain soil forms from post-depositional alteration of sedimentary parent material. In the case of alluvium, the characteristics of post-depositional alteration are useful sources of information for inferring age and soil-forming environment.

Sedimentology

Sedimentology can be considered as the encompassing study of all aspects of sedimentary deposits (Pettijohn 1975), but for this chapter we will emphasize the aspects of grain size distributions, sedimentary structures and facies assemblages that can be applied to fluvial geomorphological interpretations. The literature on sedimentological studies of ancient and modern fluvial systems is extensive. Readers interested in greater detail than provided here are referred to works by Smith and Rogers (2010) and Miall (2010).

Particle size, sedimentary structures, facies and provenance

The particle size of sediment in deposits is an indicator of the hydrological, hydraulic and sediment supply regime of the river. The size of sediment entrained or suspended can be related to calculable shear stresses, and sediment fabric (imbrication, amount of matrix) can be related to hydraulic conditions or stream power (Allen 1985; Chapters 9 and 15, this volume). The size and sorting of sediments relate in part to the size and sorting of sediment delivered to the system. The sorting and bed-scale variability of particle size may be related to hydrological variability.

Mechanical sedimentary structures – including bedding, internal bedding, bedding-plane markings and deformed bedding – are also amenable to hydraulic interpretations (Allen 1985; Gregory and Maizels 1991; Miall 2010). Bedding dimensions and grain size can be used to estimate water depth and constrain possible values of Froude number and velocity. In addition, flow-direction indicators in bedforms and internal structures can be useful in reconstructing flow patterns.

Frequently, the sedimentological information in alluvium is simplified by grouping sediments into *lithofacies* (or *facies*), units defined to have relatively uniform grain size and types of sedimentary structures (Walker and James 1992). Facies for fluvial systems have been defined based primarily on particle size and secondarily on sedimentary structures and organic content (Miall 2010). Facies can have particular genetic interpretations associated with them, indicating the type of hydraulic environment in which they form. For example, massive mud with freshwater molluscs might be interpreted as a channel-fill facies. Facies can also be aggregated into related facies assemblages in order to simplify the analysis of alluvial deposits. For example, a channel facies assemblage might be defined as a combination of massive gravel, plane-bedded gravel and trough-cross-bedded gravel facies associated with channel and point-bar environments. Alternatively, overbank and channel fill facies could be combined to define a topstratum facies assemblage (Fig. 2.1). Definitions of facies and the degree of splitting or lumping of facies assemblages ultimately depend on the utility to answer specific questions.

There are substantial practical difficulties in sampling alluvial deposits without bias or representatively for quantitative sedimentological analyses (see, for example, Wolcott and Church 1991; Rice and Church 1998). The particle size distribution of the sediment deposited by a river is a complex function of the transport capacity – as determined by available discharge and channel hydraulic conditions – and sediment availability – including quantity and degree of sorting. Transport capacity and sediment availability typically vary spatially through the channel and floodplain, creating a three-dimensional mosaic of facies through which Fig. 2.1 provides a two-dimensional vertical slice. To address a question about trends in sediment supply over time, for example, one would first have to determine age-equivalent units (sections following) and then follow one or more of the following strategies: (i) Sample the age-equivalent unit randomly in three dimensions to provide an unbiased estimate of total particle-size distribution or sedimentary features. This approach is extremely time intensive and may be practically impossible. (ii) Separate the age-equivalent units into facies units before random sampling (that is, stratified, random sampling), thereby reducing total variance and increasing efficiency. (iii) After mapping ages and facies from the best available field data, select representative samples from facies common to all units. Such samples form a basis for comparison, but would not provide unbiased estimates of mean or variance. Nevertheless, given the costs and logistic constraints on field studies, the representative sample approach is the most common in fluvial studies.

In some sedimentological studies, the provenance of the sediment may be a central question, perhaps for its value in indicating shifting sources for the sediment. Lithological, mineralogical, chemical and particle-size characteristics can be

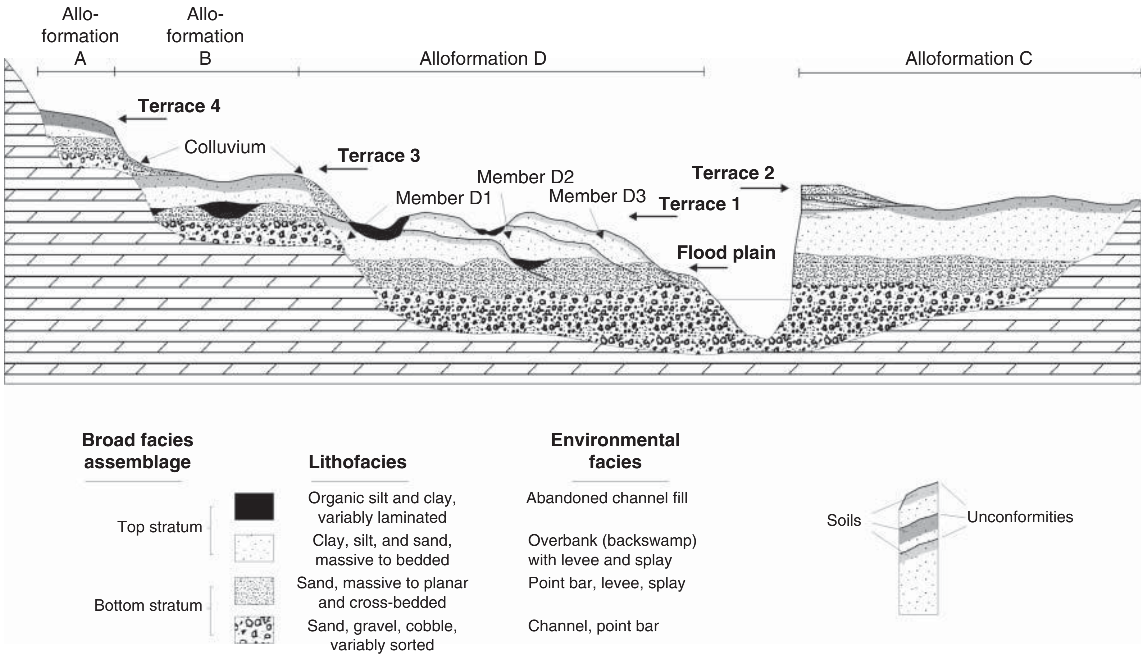


Figure 2.1 Diagrammatic cross-section of alluvial strata, showing allostratigraphic units, weathering profiles and terrace levels.

compared with distinct sediment sources to infer proportional contributions (for example, Nanson *et al.* 1995). Chemical and mineralogical characteristics of sediment are especially useful in assessing the nature of downstream sediment routing by delineating recent, contaminated sediments (for example, Magilligan 1985; Marron 1992; Owens *et al.* 1999; Malmon *et al.* 2005; Dennis *et al.* 2009).

Palaeohydraulic and hydrological interpretations

During the past few decades, stratigraphic, sedimentological and geomorphic approaches have been developed to decipher quantitatively past river flow conditions. The approaches range from empirically relating evidence of past channel morphology and facies architecture to formative flow conditions, to reconstruction of site-specific shear stresses associated with movement of an individual clast. These palaeohydraulic tools have been used to achieve understanding of a variety of geomorphic, ecological, palaeoclimatic and hazard issues.

As reviewed by Baker in 1989 and 1991, there have been three fundamental approaches to retrodict past channel and flow conditions from geological and geomorphic considerations (Table 2.1 – summary of methods, after Baker 1989). The first, termed *regime-based* (Baker 1989) relies on empirically derived relations between channel morphological or sedimentological characteristics and past flow conditions. Another classical approach to estimate past flow is through the use of *flow competence* criteria, which takes advantage of empirical and theoretical relations between some measure of flow strength and the size of clasts transported by the flow. The third approach, which has seen significant attention since the 1980s, has been the analysis

Table 2.1 Comparison of palaeohydrological and palaeohydraulic approaches.

Attribute:	Approach		
	Regime	Flow competence	Slack-water deposits
River type	Alluvial (deformable boundaries)	Alluvial and stable boundary channels	Stable boundary channels
Scale of analysis	A reach of river or channel cross-section	Individual deposit	Individual or multiple deposits within a reach of river
Commonly retrodicted properties	Mean annual discharge, 'bankfull' discharge for a channel reach or cross-section	Shear stress, velocity, unit stream power associated with individual deposit	Rare and high-magnitude floods for a channel reach
Estimated accuracy under ideal conditions	±100%	±100%	±25%
Reviews of method	Dury 1985; Williams 1988	Maizels 1983; Komar 1996	Kochel and Baker 1988; Baker 1989
Example applications	Dury 1976; Williams 1984b	Costa 1983; Williams 1983	Ely *et al.* 1993; O'Connor *et al.* 1994; Harden *et al.* 2011

of the surficial geological record of individual flood events preserved in *slack-water deposits* or other evidence of flood stage. Each of these approaches to deciphering past fluvial conditions has both unique powers and shortcomings, summarized below.

Regime-based methods of retrodicting past flow conditions have been used on a variety of alluvial systems where there is a surficial geological record of channel deposits or stratigraphic or geomorphic evidence of plan-view or cross-sectional channel geometry. These methods usually result in an estimate of high-frequency discharges (such as the mean annual flood or bankfull discharge) that are thought to control channel morphology in alluvial systems. One regime-based approach is to determine the type of channel (e.g. meandering, braided or straight) from the sedimentology of the deposits and then relate the inferred channel type to empirically defined fields of hydraulic and sediment load factors that distinguish various channel morphologies. A simple application would be to retrodict limiting channel slope and bankfull discharge conditions for deposits of either a braided or a meandering river by invoking the threshold hydraulic conditions between these two channel patterns defined by Leopold and Wolman (1957). More complex hydraulic criteria separating braided, meandering and straight channels have also been proposed (Schumm and Khan 1972; Parker 1976; Kleinhans and van den Berg 2011), which also can be used to constrain palaeohydraulic conditions such as width/depth ratio, channel slope and flow velocity. For steep alluvial streams where there is independent evidence of channel cross-section morphology, Grant's hypothesis that such streams tend to adopt cross-sectional geometries that convey flow at or near critical flow conditions (Grant 1997) can serve as a basis to retrodict velocity and discharge (Webb and Jarrett 2002).

A more commonly used regime-based approach is to use empirical relations between channel cross-section or meander dimensions and formative discharges to determine past flow conditions. Classic examples of relating meander wavelengths to palaeodischarge are provided by Dury (1954, 1965, 1976), Schumm (1967, 1968) and Knox (2006). Good examples and discussions of relating high-frequency discharge (such as bankfull or mean annual flow) to cross-section dimensions (such as depth and width) are provided by Williams (1978, 1984b) and Rotnicki (1983).

In general, regime-based retrodiction is subject to large uncertainties arising from (i) errors in assignment of the predictor variables, such as misinterpretation of channel type, cross-section shape and meander wavelength (Dury 1976; Rotnicki 1983), and (ii) the large standard errors of the empirical relations between predictor variable and discharge, which, in the most favourable cases, result in a 50% chance of error of greater than 24% (Dury 1985). Nevertheless, regime-based flow estimates can be useful for addressing questions of broad environmental change resulting in regional changes in channel behaviour. Useful and complete discussions of the various regime-based methods their uncertainties are provided by Ethridge and Schumm (1978), Rotnicki (1983), Dury (1985) and Williams (1988). Williams (1984a, 1988) provides a list of many equations used in regime-based analyses and comments on their sources and applicability.

Flow competence refers to the largest grain transported by a given discharge (Gilbert 1914). Flow strength is usually described by some measure of velocity, shear stress (force exerted by the flow parallel to the bed) or stream power (time rate of energy dissipation by the flow). Gilbert developed this concept to predict the effects of future flows, but the concept has been used since to retrodict palaeohydraulic conditions from coarse-clast deposits in the surficial geological record. Over the last 30 years, theoretical and empirical relations between clast size and flow conditions have been used to reconstruct the hydraulic conditions associated with fluvial deposits of individual flows in a wide variety of environments (Fig. 2.2),

(a)

(b)

Figure 2.2 Photographs of sites of palaeohydrological analysis. (a) Slack-water flood deposits preserved in an alcove along the Escalante River, Utah. Such deposits were used to reconstruct a history of large floods for the last 2000 years (Webb *et al.* 1988). Photograph from US Geological Survey. (b) Site of boulders deposited by the late Pleistocene Bonneville Flood near Swan Falls, Idaho. Measurements of Bonneville Flood boulder diameters were compared with reconstructed flow conditions to develop flow competence relations for large floods (O'Connor 1993). Photograph used with permission from Geological Society of America.

including Pleistocene outburst floods (Birkeland 1968; Malde 1968; Baker 1973; Lord and Kehew 1987; Kehew and Teller 1994), Holocene flood and outwash deposits (Church 1978; Bradley and Mears 1980; Costa 1983; Williams 1983; O'Connor *et al.* 1986; Waythomas *et al.* 1996; Lamb and Fonstad 2010) and Miocene turbidite deposits (Komar 1989, 1970). Palaeohydraulic studies have used relations between particle size and flow conditions proposed by Baker and Ritter (1975), Church (1978), Costa (1983), Williams (1983), Komar (1989) and O'Connor (1993). Reviews of flow competence methods, including their application and uncertainties, are provided by Maizels (1983), Williams (1983), Komar (1989, 1996), Komar and Carling (1991), Wilcock (1992), O'Connor (1993) and Lamb and Fonstad (2010).

Flow competence methods are most suitable for the reconstruction of local hydraulic conditions at the site of deposits from an individual flow. This method can be applied over a broad range of fluvial environments wherever there are coarse clastic deposits. However, because most of the empirical relations between particle size and flow strength yield predictions of local shear stress, stream power or velocity, the discharge of the flow can only be determined if there is independent information on the channel geometry. Furthermore, key assumptions must be met for a valid analysis: (i) the analysed particles must indeed have been transported and must closely represent the maximum size that could have been transported by the flow and (ii) the analysed clasts must have been transported by a water flow rather than a debris flow or other type of mass movement. Uncertainties in flow retrodiction from competence criteria are generally large and hard to quantify, primarily resulting from (i) difficulty in adequately sampling the largest particles in a deposit (Church 1978; Wilcock 1992), (ii) large standard errors associated with the empirical relations between clast size and flow conditions (Church 1978; Costa 1983) and (iii) uncertainty as to the timing of the deposit (and its retrodicted hydraulic conditions) in relation to the general inference that the deposits represent peak flow conditions (O'Connor 1993).

Slack-water deposits preserved in stratigraphic sequences along river margins have been used to provide detailed records of flood events extending back several thousand years (Baker *et al.* 1979; Patton *et al.* 1979) (Fig. 2.2). Slack-water deposits form from clay, silt and sand carried in suspension by large floods and deposited in zones of local velocity reduction. Common depositional environments include recirculation zones associated with valley constrictions or bends, tributary mouths (Baker and Kochel 1988), alcoves and caves in bedrock walls (O'Connor *et al.* 1986; O'Connor *et al.* 1994; Harden *et al.* 2011) and on top of high alluvial or bedrock surfaces that flank the channel (Ely and Baker 1985; Hosman *et al.* 2003). The sedimentary records contained in these slack-water deposits can, in certain cases, be supplemented with botanical (Hupp 1988) and erosional evidence of large floods (Ely and Baker 1985).

Most of the earliest studies of slack-water deposits were from the arid southwestern United States (Costa 1978; Baker *et al.* 1979; Kochel and Baker 1982; Kochel *et al.*1982; Patton and Dibble 1982; Ely and Baker 1985; Webb 1985), but in more recent years the scope of application has expanded to most continents (for example, Wohl 1988; Ely *et al.* 1996; Benito and Thorndycraft 2005) and into more humid environments (Knox 1988, 1993). Studies of flood stratigraphy have been motivated by questions of dam safety (for example, Ely and Baker 1985; Partridge and Baker 1987; Levish and Ostenna 1996; Hosman *et al.* 2003), climate change (Ely *et al.* 1993; Macklin and Rumbsy 2007) and geomorphic effects of floods (Webb 1985; O'Connor *et al.* 1986). Benito and O'Connor (2013) reviewed the components of a slack-water deposit study, including stratigraphic analysis and correlation, geochronology, flood discharge determination and flood frequency analysis. Detailed discussions of methods are also provided in chapters within Baker and Kochel (1988) and House *et al.* (2002). Harden *et al.* (2011) provide a recent example of how these components were woven together to provide an analysis of long-term flood frequency for streams in the Black Hills of western South Dakota, United States.

Complications of using slack-water deposits as flood-stage indicators have been documented in the case of extreme flooding in 1985 on the Cheat River in West Virginia, where measured discharges were consistently greater than calculated discharges based on slack-water stage indicators, possibly because of interactions between tributary and mainstem flow, channel instability or suspended load hysteresis resulting in little deposition at peak discharge (Kite and Linton 1991). Harvey *et al.* (2011) also suggest that slack-water depositional records may be strongly affected by watershed-scale patterns of channel erosion and deposition, thereby reducing their utility as flood recorders in certain environments.

Geochronology of alluvium

Geochronological methods for determining numerical ages for alluvial strata are numerous. Reviews of Quaternary dating methods provide comprehensive discussion (Mahaney 1984; Easterbrook 1988a; Gruen 1994). The following summary emphasizes methods applicable to the late Quaternary and typical alluvial sediments, which can be divided into three general categories: relative, numerical and hybrid (Table 2.2).

Relative methods are useful for establishing whether stratigraphic units are older or younger than others and, in some cases, can be usefully calibrated to interpolate or extrapolate ages. Relative methods based on a combination of weathering characteristics, topographic position, cross-cutting stratigraphic relations and morphology are useful for developing field criteria for regional correlations of numerically dated strata. These concepts will be discussed in more detail in following discussion of pedology and morphostratigraphy.

Numerical dating methods provide estimates of time elapsed since deposition of alluvial strata. Probably the most useful method for dating alluvium is radiocarbon dating, based on the progressive decay of ^{14}C in plant or animal material once the plant has died. With the use of accelerator mass spectrometry

Table 2.2 Geochronological methods, notes and resources.

Type	Method	Notes
Relative	Stratigraphic superposition	Highly reliable method for determining sequence of deposition. Requires good exposures or drilling, trenching observations
	Weathering characteristics	In regions with established age trends of pedogenesis, weathering rinds on clasts, desert varnish, etc., weathering characteristics can be used to determine relative age, constrain age limits and correlate units spatially (Dorn 1994; Knuepfer 1994; Pinter *et al.* 1994)
	Morphological criteria	Relative elevations of alluvial terraces can be used to determine local sequence of deposition, if there is a one-to-one relation between terrace and allostratigraphic units; if not, caution is advised. Good for regional correlations of large events, if complemented with weathering chronology and numerical dates. Degree of erosion of terraces can also be indicative of relative age (Coates 1984; Pinter *et al.* 1994)
Numerical	Radiocarbon	Most highly used radiometric method for dating alluvial sediments. Careful sampling and processing are required to reduce contamination errors. Interpretation should account for type of organic matter (that is, soil organic fractions, leaf litter, charcoal or wood) and probable effects of inherited carbon. Dendrochronologically based calibration of radiocarbon years to calendar years (Stuiver 1982) is recommended for correcting dates for secular variations in radiocarbon production (Bowman 1990; Taylor *et al.*1992) (see Chapter 9, this volume)
	Photoluminesence	Useful for dating sediments or artefacts that have been zeroed by heat or sunlight. Techniques using thermal or optically stimulated luminescence vary in precision and reliability and the techniques are evolving fast. For sediments, most reliable dating has been using loess rather than alluvium. (Gruen 1994; Duller 1996) (see Chapter 9, this volume)
	^{210}Pb	^{210}Pb generated in the atmosphere, scavenged by precipitation and adsorbed on particulates decays with a half-life of 22.3 years, providing a geochronometer for recent sediments. Calculation requires assumptions of uniform deposition rate of sediment and ^{210}Pb; the former constraint is rarely met in alluvial deposits but may be met with floodplain lakes or abandoned channels (Durham and Joshi 1984) (see Chapter 9, this volume)
	Cosmogenic isotopes	^{10}Be, ^{26}Al, ^{36}Cl, ^{3}He, ^{21}Ne, ^{14}C, ^{41}Ca cosmogenic nuclides for exposure age dating of some materials. 1 ka to 10 Ma. Requires sophisticated chemical extraction and accelerator or conventional mass spectrometry (Kurz and Brook 1994) (see Chapter 9, this volume)
	Dendrochronology	Tree rings provide detailed chronometers for dating surfaces, sedimentation rates and individual floods over short time frames (see Chapter 10, this volume)
Hybrid-correlative	Palynology	Pollen is extremely useful for developing environmental and climatic conditions and can be used by correlation to date events, such as the settlement/post-settlement boundary. Pollen is poorly preserved in many alluvial settings, however. Best results are from adjacent floodplain lakes or abandoned channels and when supplemented with radiocarbon numerical dates. For example, Royall *et al.* (1991)
	Palaeomagnetism	Magnetic properties of sediments can be used to correlate based on measures of susceptibility or remanent magnetism of sediments or heated sediments or artefacts can be compared with secular variation of the Earth's magnetic field. Correlations by secular variation have been demonstrated from <0.1 to >20 ka, but require an independently dated sequence. Works best in lacustrine depositional environments (Stupavsky and Gravenor 1984; Lund 1996)
	Archaeology	Independently dated archaeological artefacts can provide tools for relative and absolute dating and for correlation (see Chapter 3, this volume)
	Tephrochronology	Tephra found in alluvium can be correlated by chemical or petrographic techniques, combined with stratigraphic sequence, with independently dated volcanic deposits (for example, Sarna-Wojcicki *et al.* 1984, 1991)

(AMS), very small amounts of sample ($\sim$1 mg) can be used to calculate dates in the time frame 200–55,000+ years BP. Conventional radiocarbon dates (in radiocarbon years before present, BP) are generally calibrated to calendar years to account for secular variations in radiocarbon production in the atmosphere (Stuiver 1982). Secular variations in ^{14}C pose special problems for samples less than 500 years old, making it difficult to define precise age estimates for this time period. Additional errors in radiocarbon dates relate to laboratory statistical counting, sample preparation and estimates of laboratory reproducibility; these are usually reported as $\pm$ values in laboratory results. However, much greater errors can be introduced in sampling, especially in the inherited age of the carbon in the sample. A radiocarbon date from a piece of charcoal from a tree that was 800 years old when it died will overestimate the age of associated sediments by at least 800 years. Resistant materials such as charcoal and bone, in fact, may be eroded and redeposited in even younger sediments. Avoidance of these inherited age errors ('old wood errors') requires close attention to field context and sampling. One strategy is to avoid sampling resistant materials in favour of materials that would likely contribute small inherited ages, such as twigs and leaves. Radiocarbon dates should be interpreted as a maximum limiting age for the enclosing deposit. Complete discussion of assumptions and cautions of using various radiocarbon dateable materials can be found in Taylor *et al.* (1992).

In addition to the methods listed in Table 2.2, several other numerical methods deserve comment. Photoluminescence dating in alluvial sediments is based on the accumulation of a thermoluminescence (TL) or optically stimulated luminescence (OSL) signal with time after burial. TL and OSL extend to greater age than radiocarbon dating (as much as 800 ka) but with typical precisions of ±5–10% (Aitken 1997). This makes luminescence dating useful for strata that are too old or lack material for radiocarbon dating (Rittenour, 2008). The accumulation of atmospheric ^{210}Pb and ^{137}Cs also can provide a useful dating method for short time intervals if slack-water sedimentation sites are available, as described in Chapter 6.

The use of cosmogenic isotopes (^{3}He, ^{10}Be, ^{14}C, ^{21}Ne, ^{26}Al and ^{36}Cl) for exposure-age dating has been increasing dramatically in recent years (Bierman 1994). Theoretically, cosmogenic isotopes can be used to date surfaces from as little as 1 ka (^{3}He) to as much as 10 Ma (^{21}Ne) (Kurz and Brook 1994). Because exposure calculations should start with a zeroed surface or known starting inventory, most applications have been on eroded or volcanic bedrock (Weissel and Seidel 1998) or sedimentary surfaces for which inherited ages can be assumed to be small (Bierman 1994). Inherited exposures can be substantial in alluvial materials that move slowly through drainage basin, but informed selection of particle sizes can minimize inheritance errors. For example, Schmidt *et al.* (2011) found that pebbles and boulders in terrace deposits in the Andes had substantially greater inheritance than sand particles. Hallet and Putkonen (1994) discuss some the complications of applying cosmogenic surface dating to actively eroding surfaces. In cases where inheritance and erosion cannot be assumed to be negligible, constraints on erosion rates and inheritance can be incorporated in models to calculate probable age and uncertainties (Hidy *et al.* 2010).

Hybrid methods are those that can be used to estimate numerical ages through application of calibrated models. For example, calibrated models of weathering rind thickness (Lasutela *et al.* 2004), lichen growth (Matthews 1994) or desert varnish geochemistry (Dorn 1994; Schneider and Bierman 1997) can be used to estimate the ages of undated surfaces through various measurements of these properties. The presence or absence of diagnostic pollen, diagnostic tephra, macrofossils or archaeological artefacts can also provide constraints on age estimates.

Included in this hybrid group is palaeomagnetism. In the late Quaternary time frame, independently dated secular variations in magnetic field strength and orientation can provide a master curve for comparison with magnetic inclination and declination of magnetic minerals in alluvial deposits. Palaeomagnetism in this time frame is mostly performed on heated sediments, typically found in hearths buried with alluvial sediments; magnetism of heated sediments is referred to as thermal remanent magnetism (Sternberg and McGuire 1990). Remanent magnetism can also be measured from fine-grained sediments deposited in still water (detrital remanent magnetism) (Easterbrook 1988b; Verosub 1988). Such deposits, free of bioturbation, are much more likely to be found in lakes and coastal zones, but might exist in some floodplain lakes

Pedology

Pedology is the study of soil-formation processes. Physical, chemical and biological processes transform freshly deposited alluvial sediments into soil profiles with characteristics that reflect the five classic soil-forming factors: climate, topography, parent material, biological influences and time (Jenny 1941). Although complex in interaction, each of the factors is governed individually by more-or-less systematic processes. In cases where the scale and scope of study allow one or more of these factors to be considered constant (for example, climate or biological influences), the soil-forming processes can provide sufficiently systematic variations in soil profiles that the properties of the profiles can impart valuable information about geomorphic processes.

Pedogenic dating is based on the notion that variation in pedogenesis with time can be separated from the effects of other factors, allowing pedogenic characteristics to be used for relative dating, correlation and estimating deposition dates when supported by independent numerical age control (Birkeland 1984). In a dissenting opinion, Daniels and Hammer (1992) argued that it is effectively impossible to hold other factors constant – that the complexities of surface processes, different parent materials and drainage influences contribute too much variation in pedogenic characteristics to be able usefully to extract age information. Daniels and Hammer's discussion underscores the need for careful field study to assure that pedological sampling sites are chosen to minimize variation in erosional history, parent material and drainage.

In studies where the geomorphological questions being addressed are sufficiently broad, pedogenic characteristics of alluvium can be useful age indicators. For example, in arid and semi-arid areas, the accumulation of soil carbonate over time has been used very successfully in correlation and age estimation (for example, Vincent *et al.* 1994) over time frames of 10^4 years. In humid environments, accumulations of clay and iron and aluminium oxides have been found systematic over time frames of 10^4–10^7 years (Markewich and Pavich 1991).

Environmental change can also be interpreted from pedogenic characteristics. Although not as useful as pollen for general climate-change assessment, soil mineralogy can provide an integrated understanding of geochemical conditions, which can be interpreted in terms of changing temperature, drainage or water-table configuration (Sheldon and Tabor 2009). For example, micromorphological examination of concentrically zoned, secondary accumulations of Fe and Al oxides in soils have been used to document changes in soil drainage over time (Birkeland 1984; Kemp 1985).

Cumulative soil profiles provide one of the more useful pedological perspectives on processes of alluvial sedimentation, as these soils have undergone simultaneous soil formation and sediment deposition (Nikiforoff 1949). Cumulative soils can be considered part of a continuum relating degree of horizon differentiation and sedimentation rate (Fig. 2.3). Where alluvial sedimentation is rapid, periods of stable sub-aerial exposure are short or non-existent. In these cases, pedogenic alteration and bioturbation are minimal and, consequently, sedimentological information is best preserved. At the other extreme, where sedimentation rates are very slow – for example, on an alluvial terrace – pedogenic processes dominate over depositional rates and weathering information is best preserved. On low terraces and floodplains, it is common to have alternating periods of deposition and sub-aerial exposure, resulting in cumulative soil profiles. These profiles are characterized by over-thickened A horizons with high organic content and massive to weak pedogenic structure. Identification of the spatial and stratigraphic distribution of cumulative soil profiles may indicate substantial changes in river dynamics over time.

Pedofacies are laterally contiguous bodies of alluvium that differ in pedogenic attributes as a result of differing sedimentation rates (Kraus and Brown 1988). Pedofacies units (Figs 2.1 and 2.3) are distinguished by measures of pedogenesis and lithofacies. The concept can be useful in describing relative sedimentation rates of a channel and floodplain, especially in the ancient sedimentary record in aggrading environments, where relative proportions of pedofacies can be compared over long geological time intervals. Pedofacies can also be examined by statistical means, such as clustering analysis, to classify floodplains by dominant sedimentation and soil-forming processes (Bullinger-Weber and Gobat 2006).

Given the inherent complexity of soil-forming processes, pedogenic characteristics rarely provide precise indicators of age or environment. The utility of the methods ultimately

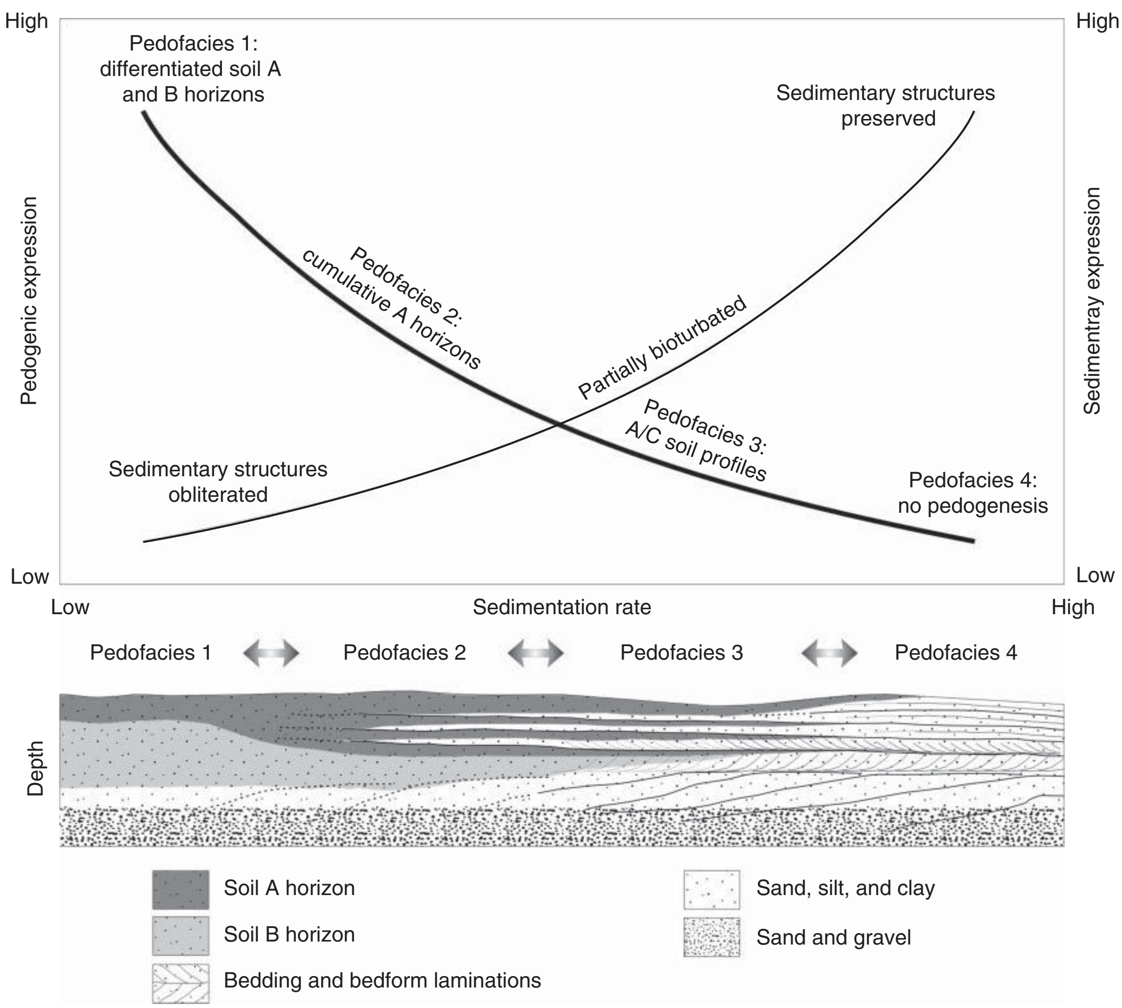


Figure 2.3 Diagram of tradeoffs between sedimentation rate and expression of pedogenic and sedimentological features.

depends, of course, on the questions being addressed. Pedogenic characteristics of alluvial strata probably never will be sufficiently precise to date high-frequency geomorphic events such as individual floods. On the other hand, pedogenic characteristics can provide useful filters for constraining geomorphic understanding and for correlation. For example, Bettis (1992) used a simple set of soil properties as regional indicators of broad age classes of Holocene alluvium and applied this filter to map the archaeological potential of alluvial deposits. Pedogenesis in this case was sufficiently systematic to sort out Early-middle Holocene, Late Holocene and Historical strata.

Stratigraphy

In this chapter, *stratigraphy* is used restrictively to denote the sequence and spatial framework of construction of the geological column (Pettijohn 1975). Stratigraphy serves to integrate sedimentology, pedology and other disciplinary approaches into a systematic understanding of how the alluvial record was constructed. Because textbooks on stratigraphy tend to emphasize long intervals of geological time and regional to continental spatial scales, they can be of limited use in geomorphological applications. Some of the best general resources in alluvial stratigraphy are in textbooks and volumes devoted to geoarchaeology (for example, Lasca and Donahue 1990; Brown 1997).

Allostratigraphic units

A useful, basic unit for describing and mapping alluvial deposits is the allostratigraphic unit, a '… mappable stratiform body of sedimentary rock that is defined and identified on the basis of its bounding discontinuities' (North American Commission on Stratigraphic Nomenclature 1983). Allostratigraphic units are similar in concept to synthems, as defined by the International Subcommission on Stratigraphic Classification, although synthems have been used to describe unconformity-bounded stratigraphic units of regional to continental scale (International Subcommission on Stratigraphic Classification 1994).

Allostratigraphic units are well suited for describing alluvial deposits because their definition allows the upper boundary to be a sub-aerial geomorphic surface (Fig. 2.1) and no constraints are put on internal characteristics, age or genesis. Hence, allostratigraphic units can be subdivided into facies, with a specified range of lithological, mineralogical, particle size or sedimentological features, age or weathering. Allostratigraphic units can be diachronous (that is, different parts of the unconformity bounded unit are of different ages) or isochronous (that is, all parts of the unit were deposited over the same time interval, within the resolution of the dating method). Allostratigraphic units are not defined by inferred time spans, but age relations may influence the choice of unit boundaries (North American Commission on Stratigraphic Nomenclature 1983). Allostratigraphic units may have pedogenic soils formed in them and soils may conform to upper and lower bounding unconformities; hence one or more pedostratigraphic units may be defined within an allostratigraphic unit and pedostratigraphic units or surface soils may be defined across several allostratigraphic units. Allostratigraphic units are usually defined as alloformations and may be aggregated into allogroups or disaggregated into allomembers (North American Commission on Stratigraphic Nomenclature 1983).

The scale of allostratigraphic units is unlimited by definition, but subject to the resolution of techniques used for measuring, tracing and mapping the units. According to the North American Code of Stratigraphic Nomenclature (North American Commission on Stratigraphic Nomenclature 1983), the only scale limitation on allostratigraphic units is that they must be mappable. Therefore, the fine-scale definition and use of alloformations may be limited by the availability of base maps and the scale used by precedent stratigraphic studies. The geographic extent of allostratigraphic units is limited by the ability to trace them continuously or to correlate from place to place based on fossil content, tephras, pedogenesis, numerical ages or topographic position.

The concept of allostratigraphic unit, therefore, provides a useful framework for describing and analysing sequences of alluvial deposits. Definition of the units is based on the bounding unconformities, so emphasis is on the sequence of erosional and depositional events; usually, these are of critical interest in geomorphological analysis. In many applications and for many alluvial deposits, it is possible and advantageous to choose alloformation boundaries based on determined ages, thereby imparting chronostratigraphic attributes to the alloformation. Differentiation of depositional lithofacies within an alloformation can be used to infer variations in depositional processes among alloformations. Differential pedogenesis of alloformations can be used to aid in the assignment of relative ages and in tracing and correlation of allostratigraphic units. Autin (1992) provided a particularly complete example of the use of alloformations in analysing the Holocene geomorphology of a large, low-gradient river in Louisiana. The general utility of allostratigraphy compared with lithostratigraphy (definition of units based on lithology) in mapping of Quaternary deposits has been the subject of some debate (Johnson *et al.* 2009; Rasanen *et al.* 2009); the arguments in favour of an allostratigraphic approach are based on its utility in defining uncomformity-bounded units relevant to alluvial histories.

Morphostratigraphy

The concept of alloformations has added useful rigour to the conventional geomorphic tool of mapping and correlating fluvial geomorphic events by the landforms they leave behind. Stratigraphic correlations can also be achieved by reference to characteristic morphology, that is, the shape and relative position of fluvial landforms. Characteristic depositional morphologies can be used to infer processes or to correlate units. For example, levee splays from a particular flood may be manifest as mappable, lobate landforms on floodplains. Morphological correlations are much stronger, however, if supported with stratigraphic, sedimentological and pedological information.

The practice of mapping and correlating terrace surfaces has underlain a great deal of geomorphological analysis, particularly at the scale of tectonic and eustatic controls on base level (Miller 1970; Bull 1991; Pazzaglia *et al.* 1998). The typical – but not universal – observation that alluvial terrace deposits at lower elevations are younger than those at higher elevations is a morphostratigraphic basis for assigning sequence and relative age (Ruhe 1975). Surface morphology also changes with age, allowing correlation based on morphostratigraphic parameters such as degree of erosional dissection and progressive erosion of depositional landforms (Pinter *et al.* 1994).

For many fluvial geomorphological studies, the question of whether alluvial units should be defined based on age, lithology, pedogenesis, sedimentology, bounding unconformities or surface morphology will depend on the questions being asked and the required resolution. The best approach for a particular study may be to use elements of multiple approaches. Hughes (2010) provides a discussion of the relative utility of different approaches to geomorphology and Quaternary stratigraphy as applied to a variety of landscapes.

Obtaining surficial geological data

Surficial geological data can only be compiled by looking into and sampling beneath the ground surface. In a typical project, the data requirements are balanced with logistical constraints of time and money and it is rare that the scientist is satisfied that all the possible relevant observations have been obtained. Several types of subsurface data can be considered. Most river reaches or segments will have natural exposures of floodplains and terraces in cutbanks. These present a low-cost but highly biased subsurface view of alluvium. Natural exposures should be observed, measured and sampled first and the knowledge gained from them should be applied to subsequent subsurface exploration. In many landscapes, man-made features such as gravel pits, pipeline trenches and road embankments also provide non-systematic opportunities for observing the subsurface.

Subsurface exploration techniques present tradeoffs that need to be considered in terms of the questions being addressed and the evolving understanding of the complexity of the alluvial deposits. Hand and power augers permit extensive probing and sampling of alluvial deposits. Fine-scale sedimentary and pedogenic features can be sampled with hydraulic split-barrel or tube samplers (Fig. 2.4). Greater depth and coarser materials require large equipment to power hollow-stem augers and even then it is rare to recover intact samples of non-cohesive sediments and it may not be possible to drill to bedrock. Shallow geophysical methods – seismic refraction/reflection, ground-penetrating radar, gravity, magnetics and electrical resistivity/conductivity methods – can be efficient means to correlate units and map the alluvium/bedrock contact, especially when geophysical data can be compared with adjacent boreholes (US Army Corps of Engineers 1998; Bersezio *et al.* 2007). No one geophysical method is applicable in all situations and different methods have different information contents (Robinson *et al.* 2008),

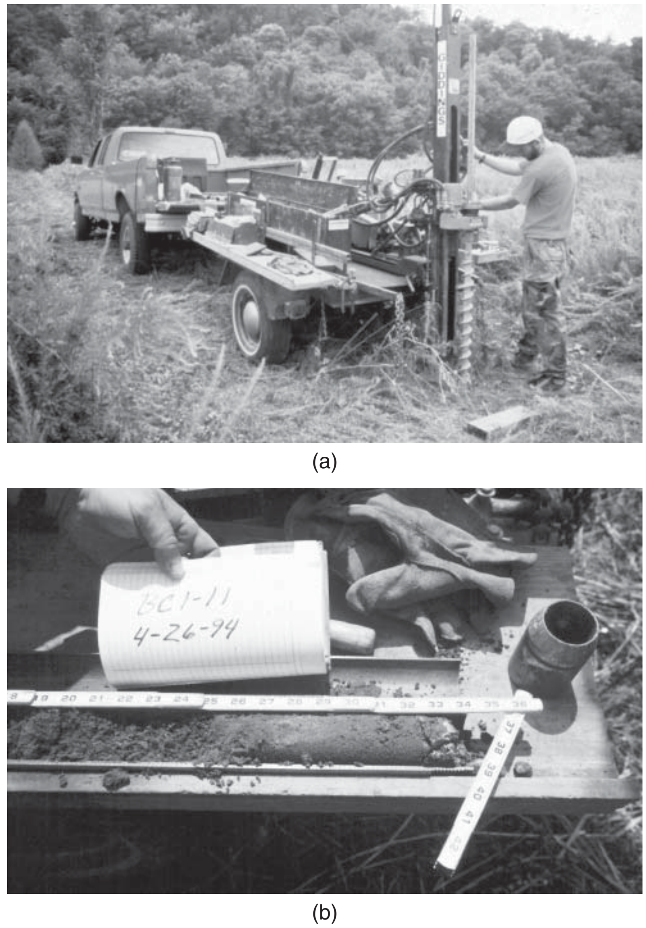


Figure 2.4 Shallow borehole drilling in valley-bottom alluvium. (a) Exploratory drilling of alluvium with a 4 in (10 cm) auger. (b) Split spoon sample of alluvium obtained by hydraulic probing. Source: US Geological Survey.

hence geophysical methods should be seen as complementary to more traditional geological approaches. Geophysical methods are developing rapidly and a complete review of applicable geophysical techniques is beyond the scope of this chapter. The interested reader is referred to texts by Sharma (1997), Gaffney (2008) and Reynolds (2011) for more extensive discussions.

Borehole logs and geophysical data arranged along surveyed topographic cross-sections may provide sufficient information to correlate units and understand the stratigraphic architecture, but boreholes generally lack the breadth of exposure needed for interpretation of many sedimentary and pedogenic features. Backhoe trenches (Figs 2.5 and 2.6) can provide long, complete exposures of near-surface strata for complete descriptions and sampling. Exposures in trenches can show meter-plus-scale bedforms, details of stratigraphic contacts and continuity of units and they provide much more efficient prospecting for datable materials or artefacts. Evaluation of soil structure, continuity of soil horizons and interpretation of environmental indicators are much more accurate in a trench than in a 1.0–2.5 cm diameter

Figure 2.5 Exploratory trench in floodplain alluvium, Big Piney Creek, Missouri. Source: US Geological Survey.

core. Placement of trenches in key places on borehole transects can improve stratigraphic understanding without requiring extensive trenching of a valley bottom. For example, trenches may be placed preferentially to sample representative features of a formation or to provide detail where contacts or facies changes occur.

Boring and augering systems presents hazards from heavy and powerful equipment. Proper personal safety equipment and kill switches are recommended and, in many localities, are required. Trenching also can present considerable hazard from cave-in. In the United States, the Occupational Safety and Health Administration, for example, requires shoring of the walls of any open trench greater than 5 ft (1.5 m) deep or stepping of the trench wall to a slope of no steeper than 34° from horizontal.

In addition, trenching and boring can create environmental hazards by delivering sediment to streams or opening up preferential pathways for contamination of shallow groundwater. These environmental hazards should be mitigated by using approved methods for filling and sealing excavations and boreholes. Many localities require permits for shallow exploratory drilling or excavations for environmental and cultural concerns. Meeting safety, environmental and cultural requirements can add considerable cost and complexity to subsurface investigations.

Geological reasoning – putting it together

Interpretation of sedimentological, geochronological, pedological and stratigraphic information can lead to geomorphic understanding if the data collection effort is carefully designed to address the question at hand and if the data are organized in a useful fashion. The task can seem daunting, but models of fluvial processes and facies architecture can help provide context. In a typical field situation, 'laws' of stratigraphic reasoning can help. Steno's law of superposition states that successively younger units overlie older units (cited in Dott and Batten 1976). Trowbridge's law of ascendancy states that terraces at higher elevations are older than those at lower elevations (cited in Ruhe 1975). Walther's law of facies states that facies that were formed in laterally adjacent environments can be found in conformable vertical sequence (cited in Reading 1978). With these concepts and good field data, lithofacies and allostratigraphic units can be assembled and put in stratigraphic sequence. Delineation of the stratigraphic units that chronicle geomorphic adjustments of a river can be accomplished best by mapping unconformity-bounded allostratigraphic units based on sedimentological and pedological characteristics.

Historically, the sequence and magnitude of fluvial geomorphic events have been inferred mainly from sequences of terrace surfaces. Such analysis is based on the assumption that depositional (cut and fill) terraces have one-to-one relations with the stratigraphic units that underlie them. Detailed stratigraphic studies of floodplains and terraces have shown that continuous terrace surfaces can overlie multiple allostratigraphic units because of onlapping or planation (Taylor and Lewin 1996). In detailed stratigraphic studies on Duck River in Tennessee, for example, Brakenridge (1984) documented that single surfaces could overlie multiple unconformity-bounded units of vertically accreted silt and clay.

The importance of delineating allostratigraphic units within terrace deposits depends on the time frame of the questions being addressed. Many recent studies have demonstrated that alluvial stratigraphic histories exhibit two dominant orders of response behaviour. Over the long term, a first-order response results in cut and fill terraces as a result of external forcing events such as climate change or base-level controls (Bull 1991). Over a shorter time frame, internal threshold responses of alluvial systems can result in second-order cut and fill sequences that may or may not form distinct topographic surfaces depending on the magnitude of the events (Schumm and Parker 1973; Bull 1991). Hence some allostratigraphic units may lack association with an external forcing event and some terrace surfaces may be underlain by multiple second-order allostratigraphic units. In addition, cut and fill stratigraphic units can form by lateral migration of a system that has surpassed intrinsic thresholds or has otherwise been unaffected by external forcing events (Ferring 1992). These units – herein called third-order cut and fill units – may be bounded by significant unconformities where the channel has migrated back into previously deposited

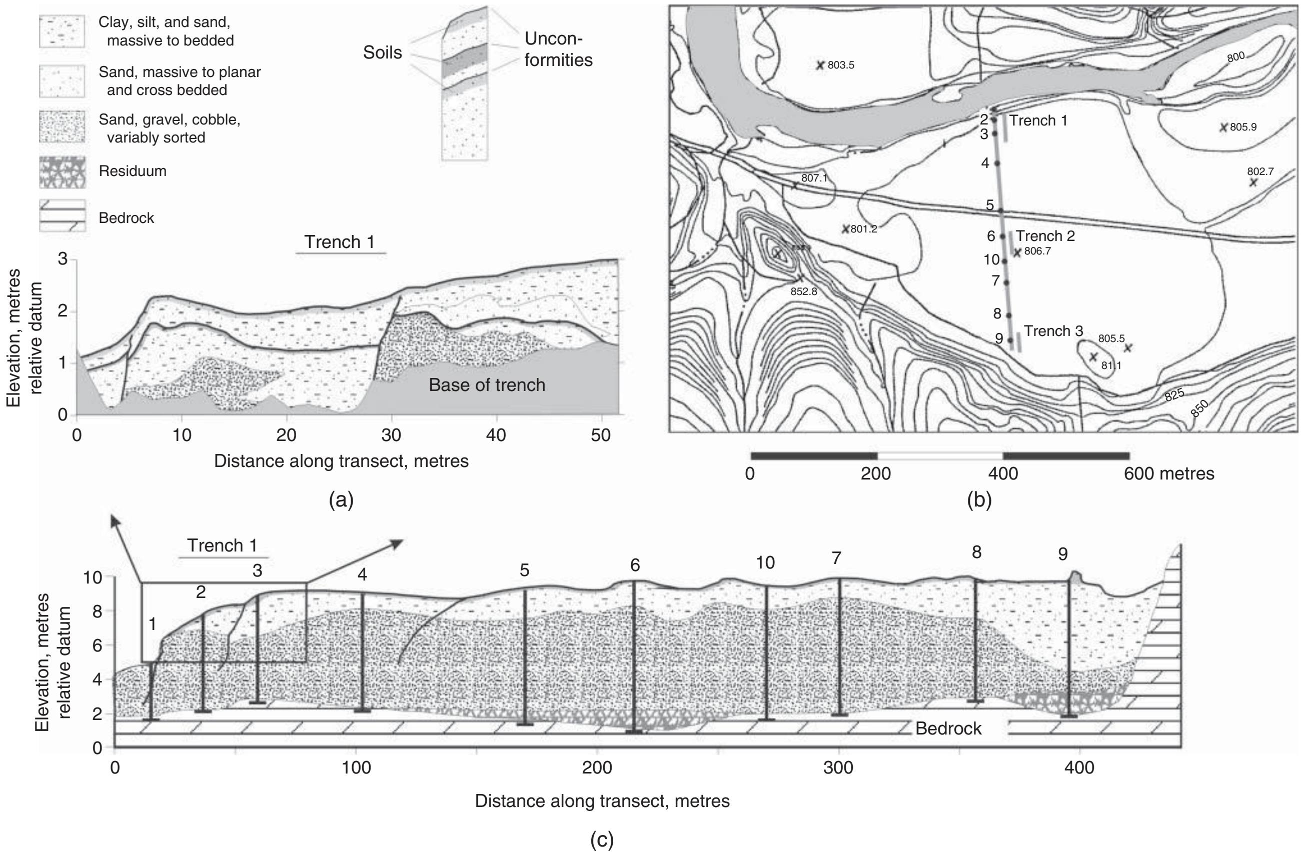


Figure 2.6 (a) Map of subsurface exploration strategy, showing locations of bore holes and trenches. (b) Close-up section of Trench 1, showing lithofacies, unconformities and buried soil profiles. This level of detail is lost in larger cross-sections compiled from bore holes, (c). Source: Albertson *et al.* (1995).

sediment, but the unconformities are not necessarily evidence of episodic forcing events.

Added to the 'noise' of second- and third-order cut and fill responses is variation in the timing of cut and fill within a drainage basin. Time lags in sediment transport in a drainage basin can result in non-synchronous deposition or so-called diachronous terrace distributions (Brown 1990; Bull 1990, 1991). For example, alluvial stratigraphic studies in smaller drainage basins in the Great Plains of the United States have shown strong correlations between moist, humid climatic conditions and aggradation and stability of Holocene deposits (Fredlund 1996). With increasing drainage area, however, the terraces become diachronous because of the lagged transport of sediment through the drainage basins and correlations with palaeoclimatic forcings diminish (Martin 1992).

Interpretations of the alluvial stratigraphy can be confounded by erosion of older units that results in gaps in the stratigraphic record. The primary determinant of the preservation potential of alluvial strata is the regional or tectonic context. Rivers in drainage basins with low sediment yield and/or low uplift rates will have low preservation potential and the alluvial record will be short and fragmented. In contrast, rivers with high sediment yield and/or rapid incision will have high preservation potential and a more complete record. Preservation occurs as downcutting and migration leave alluvial deposits behind in protected positions. Large rivers in large valleys or deltaic areas or subsiding basins will have the most complete sedimentary sequences that will tend to be preserved in the geological record, although access to these records may be more difficult because of depth of burial. Sedimentology and stratigraphy of low-gradient, aggradational rivers have been studied extensively because of their importance in the geological record (for example, Marzo and Puigdefabregas 1993). Facies and stratigraphic models developed for such rivers emphasize vertical aggradation of channel and backswamp facies over time (for example, Bridge and Mackey 1993).

In eroding river systems, the probability of preservation of a stratigraphic unit associated with an external forcing event decreases with time since deposition, but other factors also can affect preservation such as the mode of channel migration, bedrock characteristics that might shelter deposits from erosion and base-level controls. All other things being equal, the alluvial stratigraphic record will be biased toward recent and large events – so-called *preservation* or *taphonomic bias*

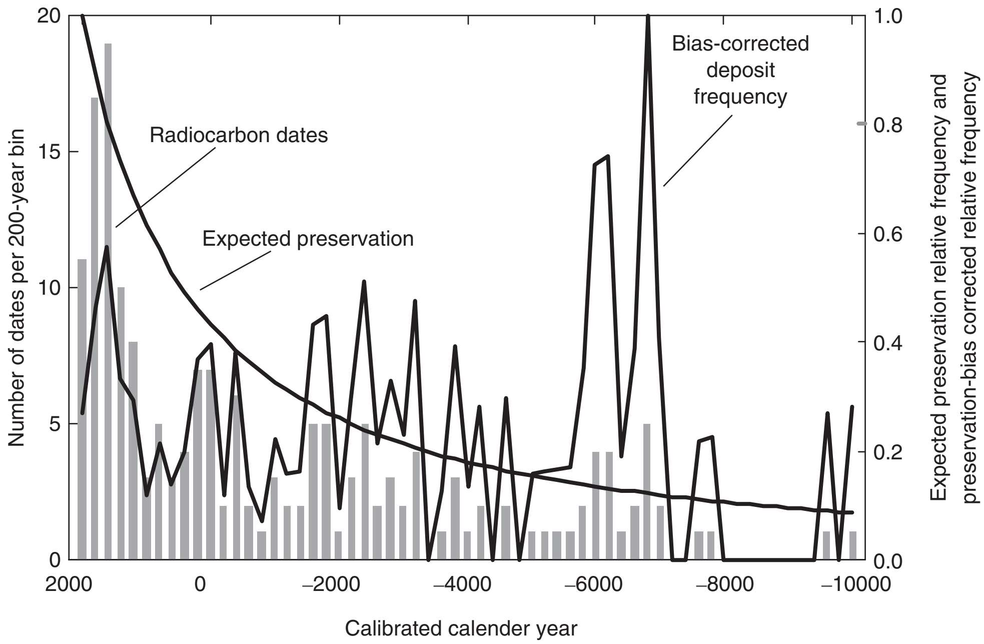


Figure 2.7 Histogram of numbers of calibrated radiocarbon dates from the Ozarks of Missouri, expected preservation model and bias-corrected deposit frequency. Data from Albertson *et al.* (1995), Haynes (1985) and R.B. Jacobson (unpublished data).

(Surovell *et al.* 2009; Ballenger and Mabry 2011). An example of preservation bias is illustrated in Fig. 2.7, which shows a frequency distribution of radiocarbon dates from the Ozarks (Haynes 1985; Albertson *et al.* 1995; R.B. Jacobson, unpublished data). The expected number of radiocarbon dated deposits was calculated from a general surficial preservation model presented by Surovell *et al.* (2009) and is shown for reference, although a model specific to the Ozarks – if it could be independently established – would be preferred. By correcting to the expected relative frequency of preservation, it becomes clear that a large number of deposits in 1400–1800 AD likely result from inherent bias in preserving young deposits; a prominent departure of the corrected data for 6000–7000 BC is more likely the result of a forcing event. Another useful approach to evaluating large populations of radiocarbon-dated sediments is to sum the calibrated probability distributions before normalizing for preservation bias to provide a continuous model of depositional events (Lewin *et al.* 2005; Harden *et al.* 2010; Macklin et al 2010).

Preservation can also be biased by type of depositional event and by geological controls. In the case of extreme flood events, if a river does not migrate actively and create new floodplains, the record of floods is progressively filtered because only sediments from larger and larger floods are preserved (Wells 1990). Moreover, narrow bedrock canyons preserve mainly deposits of infrequent floods – as there is little accommodation space for preservation of intermediate flood events – whereas wide alluvial reaches can preserve deposits from a wider range of flow events (Harden *et al.* 2010; Harvey and Pederson 2011).

2.3 Applications of surficial geological approaches to geomorphic interpretation

In this section, we present examples of how surficial geological tools have been applied to some geomorphic problems. Our emphasis is on illustrating the use of surficial geological tools rather than completely reviewing the field.

Palaeohydrological interpretations from surficial geological data

Surficial geological investigations of alluvium in Japan demonstrate how the stratigraphic record can be used to evaluate the sensitivity of the landscape to climate change and to gain insight into long-term flood frequency. Systematic changes in gravel facies in Japanese alluvial fans have been related to climatic change. At present, the Japanese Islands are characterized by frequent heavy storms. The daily maximum rainfall record exceeds 300 mm for most of Japan, among the world's highest (Matsumoto 1993). About every 10 years hourly rainfall exceeds 50 mm (Iwai and Ishiguro 1970), triggering widespread slope failure in mountainous areas (Oguchi 1996). Heavy rains occur during the typhoon season (mostly August–October) and during the Japanese rainy season (June–July) when the Polar front stays over Japan. The frequent storms lead to an abundant supply of clastic materials from mountain slopes, rapidly transported to piedmont areas. Consequently, alluvial fans are abundant in Japan: 490 alluvial fans each with an area of more than 2 km^2 occur within the Japanese Islands (Saito 1988).

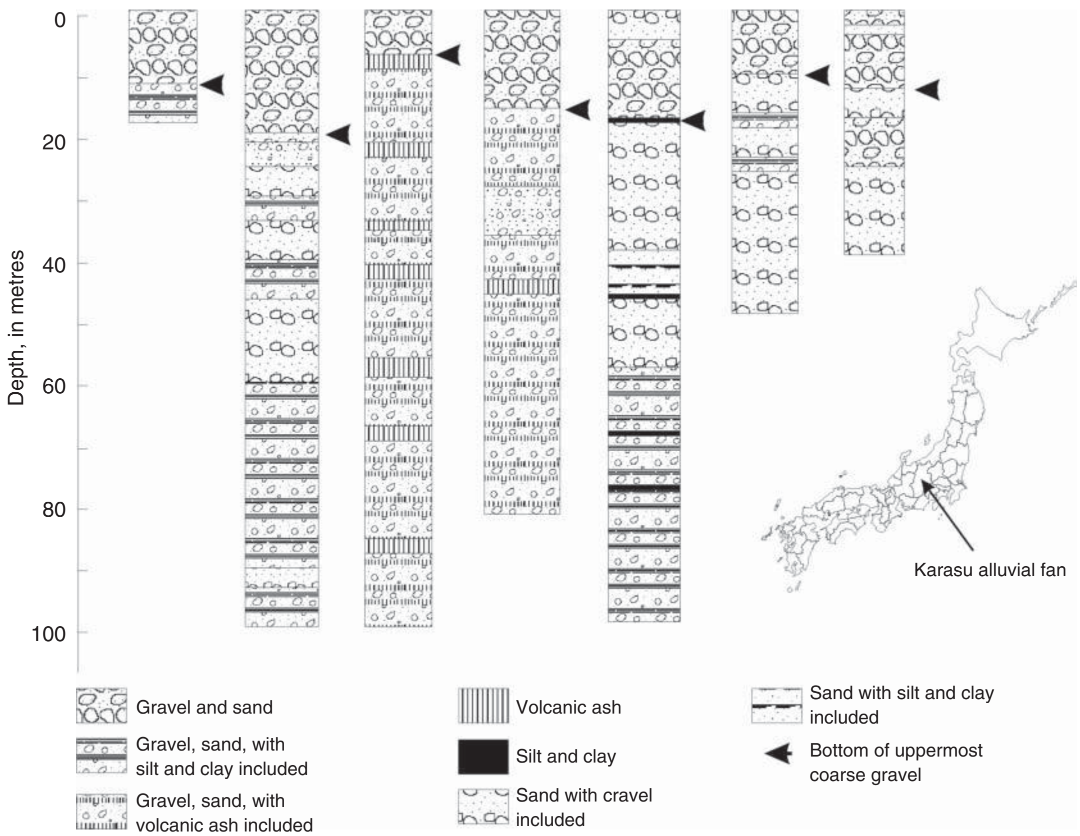


Figure 2.8 Columnar sections of alluvial fan deposits along the Karasu River, Japan. Source: Oguchi *et al*, 1997. Reproduced with permission of Journal of Quaternary Science.

Surficial geological studies of Japanese alluvial fans indicate that Holocene climatic conditions are substantially different from Pleistocene conditions. An extensive investigation of the 490 large alluvial fans in Japan revealed that particle size distributions of alluvial fan deposits dating from the Last Glacial Maximum are generally finer than those of Holocene deposits (Saito 1988). Borehole data (Fig. 2.8) at the Karasu Alluvial Fan in an intermontane basin of central Japan show how gravel sizes differ between the Last Glacial and post-glacial time. The fan deposits have been supplied from the Northern Japan Alps, which consist mostly of steep hillslopes with a modal angle of about 35° (Katsube and Oguchi 1999). The Holocene fan deposits include abundant coarse gravel with sandy matrix, reflecting the fact that about 80% of contemporary alluvial fan sediments in mountain areas of Japan are transported as bed load (Oguchi 1997). By contrast, the Last Glacial deposits are characterized by finer matrix including silt and clay as well as smaller gravel sizes.

The contrasting lithofacies of these units is used to deduce the effect of the Pleistocene–Holocene climatic transition in Japan. Around the Last Glacial Maximum, the southward shift of frontal zones led to significantly reduced storm intensity in Japan, because both typhoons and the Polar front did not attack the Islands (Suzuki 1971). The decrease in heavy rainfall resulted in smaller sediment supply from hillslopes, lower tractive force of stream flow and reduced sizes of transported gravel, compared with the Holocene (Sugai 1993). The marked change in gravel sizes also is useful for estimating the rate of post-glacial sedimentation at alluvial fans. Subsurface contours representing the boundary between the Last Glacial and post-glacial fan deposits have been drawn for the eastern foot of the Japan Alps using data from approximately 120 boreholes (Tokyo Bureau of International Trade and Industry 1984; Oguchi 1997). The volume between the boundary and the present Earth surface for each alluvial fan can be computed to estimate post-glacial sediment storage. The volumetric comparison between the storage and inferred sediment supply from upstream areas (Oguchi 1997) suggests that a significant portion of the sediment supplied during the Holocene has been stored in the alluvial fans. This is due to a large percentage of coarse bedload in post-glacial fan sediments, which are not easily transported downstream from alluvial fans.

Although clear stratigraphic evidence of slack-water deposits is thought to be rare in humid regions because of disturbance by bioturbation and pedogenesis (Kochel *et al.* 1982; Baker 1987), a study of the Nakagawa River in central Japan revealed that well-preserved Holocene slack-water deposits can occur in a humid region with abundant rainfall (Jones *et al.* 2001). The field section of the deposits is exposed on the outer bend of a meander in a gorge that cuts into late Pleistocene river

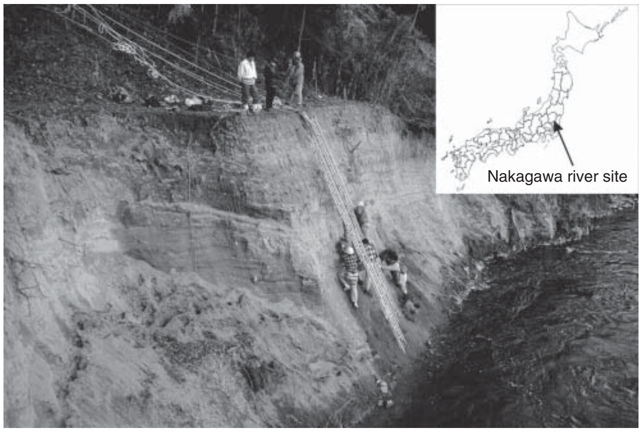


Figure 2.9 Photograph of the field section on Nakagawa River, showing bedsets and laminasets used for reconstructing flood history. Photograph: T. Oguchi.

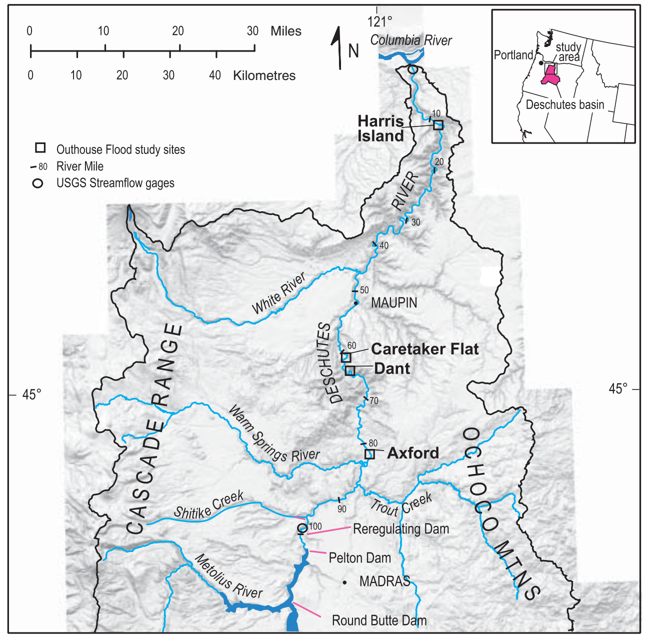


Figure 2.10 Location map showing lower Deschutes River and key Outhouse Flood study sites along the lower 130 km of river corridor. Source: Hosman *et al.* 2003. Reproduced with permission of American Geophysical Union.

terraces. The section is about 25 m in length and 8 m in height (Fig. 2.9). The sediments consist mainly of sand with numerous fine laminations and thin beds, although gravel units occur intermittently throughout the section. Radiocarbon dating and sedimentological analyses indicate that the deposits were accumulated by more than 30–40 flood events during the last 500 years. The inferred recurrence interval of palaeofloods is much shorter than that in arid to semi-arid regions and the sedimentation rate of the deposits is much higher, which can be explained by the frequency of large storms and their associated sediment loads. Indeed, three major floods in 1986, 1992 and 1998 caused repeated riverside sedimentation in the watershed of the Nakagawa River. Despite the possibility for rapid bioturbation and pedogenesis under a humid climate, their effects are limited at the Nakagawa section because of very fast and frequent sedimentation.

Catastrophic events: exceptional floods and channel and valley-bottom morphology on the Deschutes River, Oregon

The Deschutes River in central Oregon drains 28,000 km^2 of north-central Oregon, joining the Columbia River 160 km east of Portland (Fig. 2.10). Three hydroelectric dams impound the river 160–180 km upstream from the Columbia confluence and the effects of these dams on channel geomorphology and aquatic habitat have been studied by McClure *et al.* (1997), Fassnacht (1998), Fassnacht *et al.* (1998), McClure (1998) and O'Connor *et al.* (1998); most of this work is summarized in O'Connor and Grant (2003). There are few clear effects on the channel and valley bottom that can linked to the nearly 50 years of impoundment because: (i) there has been little alteration of the hydrological regime; (ii) sediment yield from the catchment is low, so the effect of trapping sediment behind the dams is less here than elsewhere; and (iii) much of the present channel and valley bottom has been shaped by exceptional floods much larger than the largest historic meteorological floods of 1964 and 1996. This section summarizes studies (Beebee and O'Connor, 2003; Hosman, *et al.* 2003) of the 'Outhouse Flood' – a large, late Holocene flood on whose deposits numerous campsite outhouses (pit toilets) have been built.

Evidence for the Outhouse Flood, a Holocene flood (or floods) much larger than the largest gauged floods of December 1964 and February 1996, includes high cobble and boulder bars at several locations along the 160 km length of the Deschutes River canyon between the dam complex and the Columbia River. The bar forms left by the Outhouse Flood and their relation to maximum stages of the February 1996 flood are particularly clear at Harris Island at River Mile (RM) 11 (Fig. 2.11) where the tops of the cobbly bar crests are 5–6 m above summer water levels and 1–2 m above the highest February 1996 inundation. Outhouse Flood bars and trimlines are 1–7 m higher than February 1996 flood stages at many other locations also (Fig. 2.12). The positions of coarse Outhouse Flood deposits along the inside of canyon bends, at canyon expansions and upstream of tight canyon constrictions are as would be expected considering the hydraulics of a large flow occupying the entire valley bottom (for example, Bretz 1928; Malde 1968). The rounded morphologies with boulder-covered surfaces that rise in the downstream direction rule out the possibility that they are terraces.

The age of the Outhouse Flood is only loosely constrained. Outhouse Flood deposits sampled from a backhoe trench at Harris Island (Fig. 2.13) contain pumice grains from the 7700 BP (calendar years) eruption from Mt Mazama. Likely Outhouse Flood deposits at RM 62 are stratigraphically below a hearth which yielded a radiocarbon age of 2910 ± 50 ^{14}C years

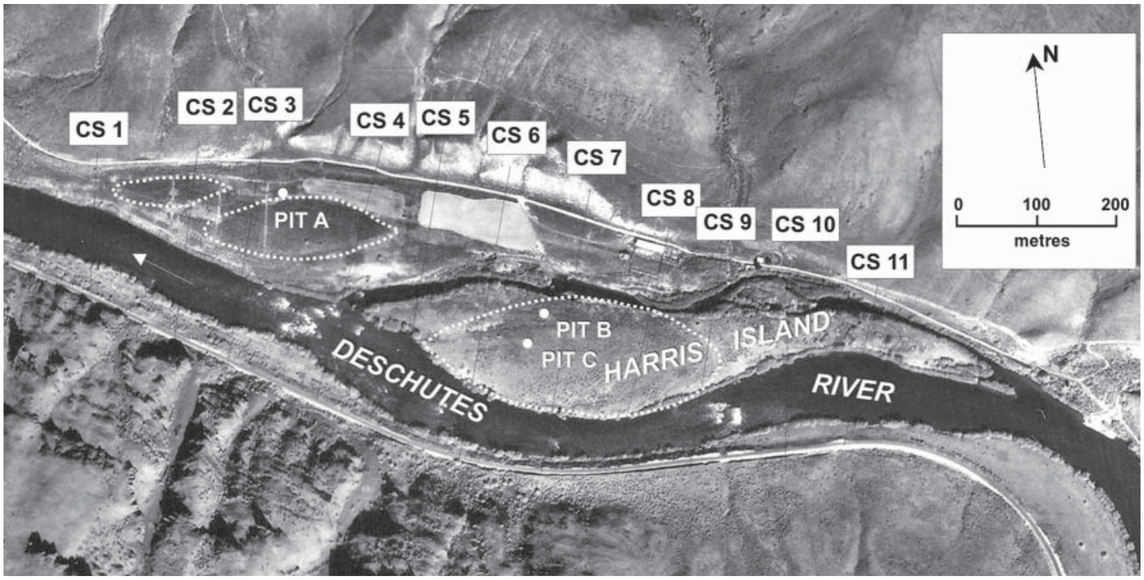


Figure 2.11 Portion of a vertical aerial photograph (WAC-95OR; 10–85; 28 March 1995) of the area around Harris Island (RM 12), showing surveyed cross-sections, locations of trenches and outlines of three flood bars that formed in this valley expansion.

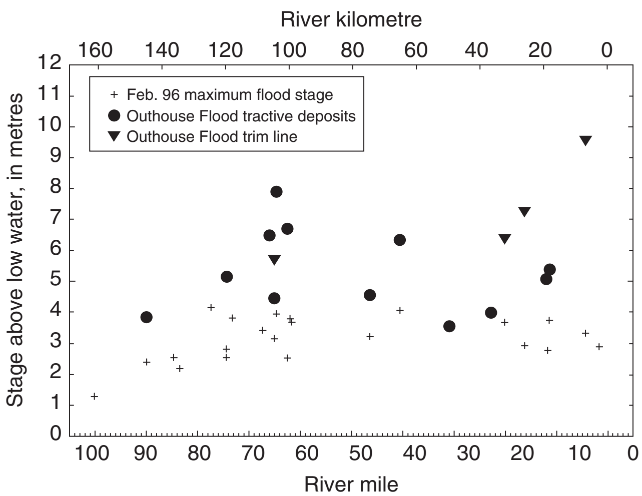


Figure 2.12 Maximum elevation of the February 1996 flood, Outhouse Flood deposits and prominent trim lines formed in Pleistocene alluvial deposits above Outhouse Flood bars. Elevations are referenced to river level, which varied less than 0.3 m during the times of surveys. Surveys were conducted by tape and inclinometer. Also included are stages of February 1996 flood recorded at US Geological Survey gauges 'Deschutes River near Madras' (Station 14092500, River Mile 100.1) and 'Deschutes River at Moody, near Biggs, Oregon' (Station 14103000, River Mile 1.4).

Figure 2.13 Photograph of backhoe trench excavated into Outhouse Flood bar at Harris Island (Deschutes River Mile 11; pit C of Figure. 2.11). The deposit is composed of rounded to subrounded basalt clasts stratified into subhorizontal, clast-supported layers distinguishable by variations in maximum clast size. Pumice grains collected from the sandy deposit matrix match tephra produced by the 7700 calendar year BP eruption of Mt Mazama (Andrei Sarna-Wojcicki, US Geological Survey, personal communication 1999), indicating that the deposits are younger than 7700 years BP. Gradations on the stadia rod are 0.3 m (1 ft). Source: US Geological Survey.

BP (Beta-131837, equivalent to 1220–1030 BC in calibrated calendar years) (Stuiver and Kra 1986).

Constraints on the peak discharge for this flood were estimated using the Manning's equation at a surveyed cross-section at Harris Island (Fig. 2.14). An n value was selected to match gauged discharge of the 8 February 1996 flood. The top of the flood bar at Harris Island requires that the Outhouse Flood had a maximum stage of at least 5.5 m above the summer low water surface. Assuming the present valley and channel bottom geometry and the slope and roughness parameters noted above, the discharge of the Outhouse Flood exceeded 5000 $m^3 s^{-1}$. A more realistic discharge estimate of 12,500 $m^3 s^{-1}$ is obtained by assuming that the water surface was about 2.5 m higher than the top of the bar and achieved a maximum stage of about 8 m above the summer water surface – a value consistent with the bouldery composition of the tops of the bars and the elevations of trim lines upstream and downstream of Harris Island (Fig. 2.12).

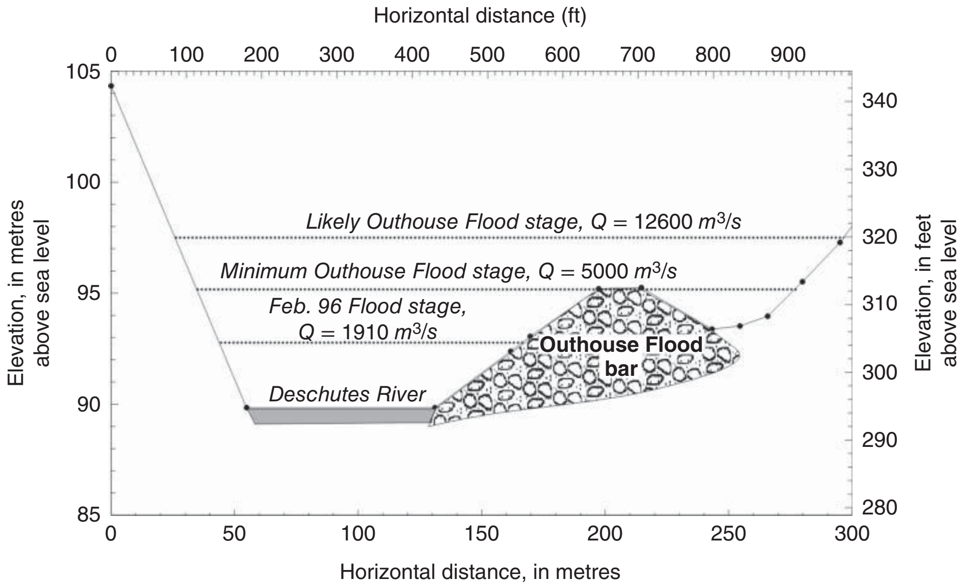


Figure 2.14 Cross-section and stages used for Manning's equation estimates of the discharge of the Outhouse Flood at Harris Island. The 'likely Outhouse Flood stage' was estimated from local elevation of prominent trim lines above nearby Outhouse Flood bars (Figure 2.13). Also shown is the maximum stage and discharge for the February 1996 flood, which was gauged at 1910 $m^3 s^{-1}$ at the Moddy Gage 19 km downstream. Outhouse Flood discharges were calculated using a measured reach-scale slope of 0.02 and a Manning's *n* value of 0.045, which was derived based on the known stage and discharge of the February 1996 flood. The cross-section corresponds to cross-section CS 3 of Figure 2.11.

These calculations can be considered conservative because they assume no valley or channel scour and use a relatively large Manning's *n* value (0.045). These estimates indicate that the Outhouse Flood was 2.5–5 times larger than the largest historic flow recorded in nearly 100 years of record.

We have no evidence for the source of the Outhouse Flood, but the distribution of similar high boulder deposits along the entire Deschutes River canyon below the Pelton Round Butte dam complex leads us to conclude that the flood came from upstream of the complex rather than from a landslide breach or some other impoundment within the canyon. The exceptionally large discharge seems greater than could plausibly result from a meteorological event like the 1964 and 1996 floods, although we cannot rule out that possibility.

The effects of the Outhouse Flood on the present valley bottom are clear and substantial. About 35% of the valley bottom between the dam complex and the Columbia River confluence is composed of cobbly and bouldery alluvium interpreted to have been deposited by the Outhouse Flood. Additionally, the five largest islands in the Deschutes River downstream of the dam complex are large mid-channel flood bars left by this one ancient flood.

The bars deposited by the Outhouse Flood have left a lasting legacy that is relevant to assessing effects of the dam on the channel. The clasts composing these large bars are larger than can be carried by modern floods and large portions of these bars stand above maximum modern flood stages. Only locally are these large bars eroded where main flow threads attack bar edges, but nowhere does it appear that cumulative erosion has exceeded more than a few per cent of their original extent. Consequently, for many locations along the Deschutes River valley, the present channel is essentially locked into its position by the coarse bars. Modern processes – associated with pre-dam or post-dam conditions – are unable to modify the valley-bottom morphology substantially. This case study emphasizes the importance of understanding the surficial geological and palaeohydrological context of individual river systems before can one fully assess the potential of environmental stresses to cause changes in channel or valley-bottom conditions.

Land-use effects and river restoration

Land-use changes can affect runoff, sediment supply, sedimentation or all of these factors, resulting in extensive changes to rivers and the ecosystems they support (for example, Wolman 1967; Trimble 1974; Arnold *et al.* 1982; Trimble and Lund 1982; Nolan *et al.* 1995; Collier *et al.* 1996; Walters and Merritts 2008). Restoration of a river requires two critical concepts: a reference condition to define restoration goals and a process-based understanding of how to attain the goals. Information in the surficial geological record can be used to construct both of these concepts. The surficial geological record is sometimes the only source of information to describe the natural, pre-disturbance condition in highly disturbed river basins. In particular, the surficial geological record is a critical source of information for determining whether restoration is necessary by providing a long-term record of natural variation. The surficial geological record can also be used to diagnose what has happened to

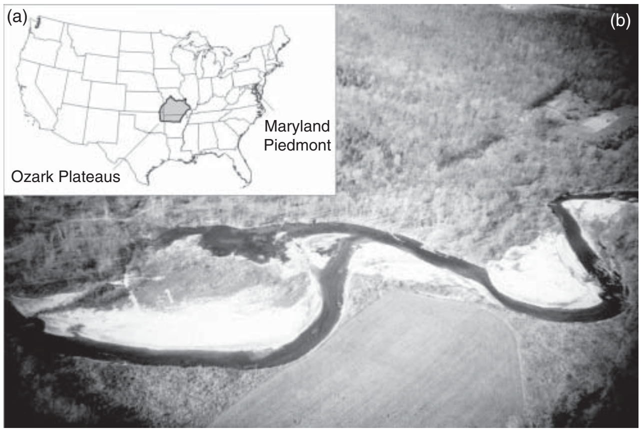


Figure 2.15 (a) Location map of land-use effects examples. (b) Photograph of a typical Ozarks stream, with extensive, unstable gravel bars. Source: US Geological Survey.

degrade a river system and thereby to develop an understanding of how to restore it.

In the Ozarks of Missouri (Fig. 2.15), for example, there has been a pervasive belief that streams have too much gravel in them, indicated by the large number of extensive, unstable gravel bars. The abundance of gravel has been attributed to massive erosion associated with timber harvest in the period 1880–1920 (Kohler 1984; Love 1990). Surficial geological investigations of valley bottoms have provided a better understanding of gravel in Ozarks streams and how streams have responded to land-use changes (Table 2.3) (Jacobson and Pugh 1992; Albertson *et al.* 1995; Jacobson and Primm 1997). These investigations have documented that large quantities of gravel were deposited in Ozarks streams throughout the Holocene (Fig. 2.16a). This context indicates that present-day gravel distributions are not extreme aberrations. Nonetheless, stratigraphic sections indicate that in 4–5th-order streams, greater quantities of gravel have been deposited over the last 60–130 years than previously, an observation that confirms popular perceptions that the streams have been quantitatively altered by land-use changes. A more dramatic and unexpected effect, however, has been a decreased deposition of fine sediment (silt and clay) over the same time interval. This observation has focused attention on the role of riparian land use and vegetation in providing hydraulic roughness and consequent deposition of fine sediments (McKenney *et al.* 1995). The dominant mode of aggradation of land use-derived gravel has been the lateral accumulation of extensive inset point bars with greater thicknesses than before settlement. Lateral aggradation of coarse sediment is favoured in these watersheds because of the great quantities of chert produced by weathering of Palaeozoic carbonate rocks.

The stratigraphic history of sedimentation in Ozarks streams has been an integral part of studies linking land-use changes to stream habitats and ecological processes (Peterson 1996; Jacobson and Primm 1997; Jacobson and Pugh 1997; McKenney 1997; Jacobson and Gran 1999; Jacobson, 2004). Surficial geological tools were especially important for providing a qualitative understanding of what Ozarks streams looked like prior to European settlement and for identifying changes in channel processes.

The Ozarks example of gravel aggradation provides a useful comparison with other studies of land use-induced aggradation. Most documented stream responses to agricultural land-use changes in the humid, eastern half of the United States have been dominated by aggradation of fine sediment (for example, Trimble 1974; Costa 1975). Jacobson and Coleman (1986) documented vertical aggradation of floodplains in several stream valleys in the Maryland Piedmont (Fig. 2.16b). Vertical aggradation of overbank sediments was dominant because of the abundance of fine sediment produced by weathering of metasedimentary rocks in these watersheds, because of proportionately greater increases in sediment supply compared with increases in runoff for a given increase in agricultural land use and because many of the valley bottoms were occupied by mill ponds that served to retain sediment (Walters and Merrits 2008). In the past 60 years or so, vertical aggradation has been replaced by lateral aggradation of sand and gravel as soil-erosion controls have decreased fine sediment loads. In addition to focusing attention on the role of sediment supply in basin instability, the alluvial stratigraphic record in the Piedmont indicated which valley-bottom surfaces were appropriate to use for measuring bankfull channel dimensions (Coleman 1982). Moreover, the surficial geological history documented the large quantity of floodplain sediment that could be remobilized and delivered rapidly to streams as a result of lateral migration.

Table 2.3 Alloformations defined for south-central Ozarks alluvial deposits.

Alloformation	Lithological and pedological features	Age range (calibrated calendar years)
Cookesville	Actively aggrading point bars, alternate bars and channel; sand, gravel and cobbles; negligible pedogenesis	Present–1850 AD
Happy Hollow	Stratified sand and gravel on low valley-bottom surfaces; very weak soil structure; multiple A–C horizons	Present–1650 AD
Ramsey	Sandy silt overbank and gravel bottom stratum; cambic B horizons. A–Bw–C1 soil profiles	1650–550 AD
Dundas	Loamy overbank and gravel bottom stratum; weak argillic B horizons, partly oxidized with 7.5YR4/4–5/4 colours. Ap-Bw–Bt–C soil profiles	50 AD–1050 BC
Miller	Silty, thick overbank deposits and gravel bottom stratum; moderate to well-developed soil structure with oxidized B horizons with 7.5YR5/6 colours. Ap–Bt–C soil profiles	2350–8050 BC

From Albertson *et al.* (1995).

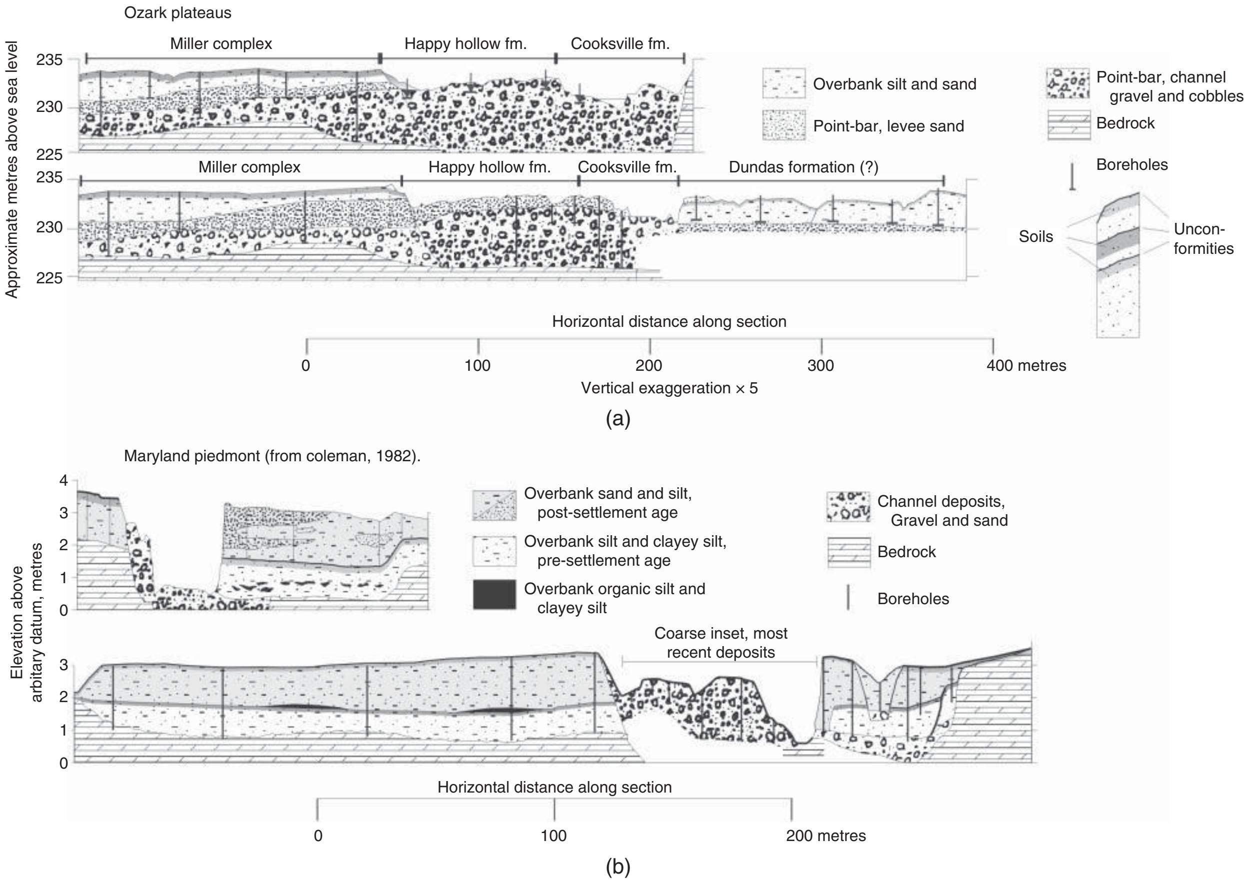


Figure 2.16 Stratigraphic cross-sections. (a) From Missouri Ozarks, showing post-settlement aggradation of coarse-grain, point-bar and channel facies and loss of fine, overbank facies. Land use-induced aggradation is shown to be dominantly lateral accumulation of coarse-grained Cooksville and Happy Hollow alloformations, Table 2.3. (b) From Maryland Piedmont, showing post-settlement, vertical floodplain aggradation by fine, overbank and millpond facies. Vertical floodplain aggradation was succeeded by deposition of coarse channel and floodplain deposits when the sediment supply was decreased (Jacobson and Coleman 1986). Source: Coleman *et al*, 1982. Reproduced with permission of Derrick Coleman, John Hopkins University.

2.4 Summary and conclusions

Surficial geological tools fill an important role in fluvial geomorphic studies. Sedimentology, geochronology, pedology and stratigraphy in combination can extend the record of river dynamics and provide essential context for predictive understanding. The chronicle of geomorphic changes preserved in alluvium, although subject to gaps and requiring interpretation, is primary evidence for how a river system has responded to past environmental stresses. Understanding of past dynamics improves prospects for assessing the present state of the river and for constraining predictions of future behaviour.

A holistic understanding of alluvial deposits requires the application of many disciplines, each of which has substantial complexity of its own. One of the most critical decisions in the application of surficial geological tools is how to limit a study, to collect data that are most efficient in addressing the geomorphological question at hand. Some geomorphological questions can involve relatively simple approaches. For example, creation of a flood hazard map may require simply mapping out Recent and Late Holocene deposits using hand augers and lumping the remainder of valley bottom deposits into a low-hazard category. In this situation, a small investment in surficial geology could produce substantial increases in understanding. In contrast, palaeohydrological and palaeohydraulic studies to reconstruct the effects of climate change over the last 15,000 years would require a great deal more information from a similar floodplain. Such investigation could include detailed lithofacies maps, detailed definitions of allostratigraphic units, palaeochannel and palaeohydraulic reconstructions, numerical dates from multiple sources and environmental indicators of climatic conditions. In this type of situation, a substantial investment of time and effort could yield large quantities of detailed information on the timing and magnitude of geomorphological changes.

Inevitably, however, there must be diminishing marginal returns on the investment in surficial geological data. At some point, the geologist runs into the holes between the sediment (Ager 1993) or finds that past conditions are inadequate analogues for future conditions. At this point, the geomorphologist must turn to more complete sedimentary records (such as lakes or oceans) or other analytical tools to develop predictive understanding. In our experience, we have found that predictive tools have much greater utility when they are selected and calibrated based on the surficial geological record of the past.

References

Ager, D.V. 1993. *The Nature of the Stratigraphical Record*, Chichester: John Wiley & Sons, 151pp.

Aitken, M.J. 1997. Luminescence dating. In: Taylor, R.E. and Aitken, M.J., eds., *Chronometric Dating in Archaeology*, New York: Plenum Press, pp. 183–216.

Albertson, P.E., Meinert, D. and Butler, G. 1995. Geomorphic Evaluation of Fort Leonard Wood, *Technical Report GL-95-19*, Vicksburg, MS: US Army Corps of Engineers, Waterways Experiment Station.

Allen, J.R.L. 1985. *Principles of Physical Sedimentology*, London: George Allen and Unwin, 272pp.

Arnold, C.A., Boison, P.J. and Patton, P.C. 1982. Sawmill Brook: an example of rapid geomorphic change related to urbanization. *Journal of Geology* 90: 115–166.

Autin, W.J. 1992. Use of alloformations for definition of Holocene meander belts in the middle Amite River, southeastern Louisiana. *Geological Society of America Bulletin* 104: 233–241.

Baker, V.R. 1973. Paleohydrology and Sedimentology of Lake Missoula Flooding in Eastern Washington, *Special Paper 144*, Boulder, CO: Geological Society of America, 79pp.

Baker, V.R. 1977. Stream-channel responses to floods, with examples from central Texas. *Geological Society of America Bulletin* 88: 1057–1071.

Baker, V.R. 1987. Paleoflood hydrology and extraordinary flood events. *Journal of Hydrology* 96: 79–99.

Baker, V.R. 1989. Magnitude and frequency of palaeofloods. In: Beven, K. and Carling, P., eds., *Floods – Hydrological,* Sedimentological and Geomorphological Implications, Chichester: John Wiley & Sons, pp. 171–183.

Baker, V.R. 1991. A bright future for old flows. In: Starkel, L., Gregory, K.J. and Thornes, J.B., eds., *Temperate Palaeohydrology*, New York: John Wiley & Sons, pp. 497–520.

Baker, V.R. 1996. Discovering Earth's future in its past: palaeohydrology and global environmental change. In: Branson, J., Brown A.G. and Gregory K.J., eds., *Global Continental Changes: the Context of Palaeohydrology*, London: Geological Society, pp. 73–83.

Baker, V.R. and Kochel, R.C. 1988. Paleoflood analysis using slackwater deposits. In: Baker, V.R., Kochel, R.C. and Patton, P.C., eds., *Flood Geomorphology*, New York: John Wiley & Sons, pp. 357–376.

Baker, V.R. and Ritter, D.F. 1975. Competence of rivers to transport coarse bed load material. *Geological Society of America Bulletin* 86: 975–978.

Baker, V.R., Kochel, R.C. and Patton, P.C. 1979. Long-term flood frequency analysis using geological data. In: *The Hydrology of Areas of Low Precipitation: Proceedings of the Canberra Symposium, December 1979*, IAHS–AISH Publication 128, Louvain: International Association of Hydrological Sciences, pp. 3–9.

Ballenger, J.A.M. and Mabry, J.B. 2011. Temporal frequency distributions of alluvium in the American Southwest: taphonomic, paleohydraulic and demographic implications. *Journal of Archaeological Science* 38: 1314–1325.

Beebee, R.A. and O'Connor, J.E. 2003. The Outhouse Flood: a large Holocene flood on the lower Deschutes River, Oregon. In: O'Connor, J.E. and Grant, G.E., eds., A Peculiar River: Geology, Geomorphology and Hydrology of the Deschutes River, Oregon, *Water Science and Application Series*, vol. 7, Washington, DC: American Geophysical Union, pp. 147–168.

Benito, G. and O'Connor, J.E. 2013. Quantitative paleoflood hydrology. In: Wohl, E.E., ed., *Fluvial Geomorphology. Treatise on Geomorphology*, vol. 9, San Diego, CA: Academic Press, pp. 459–474.

Benito G. and Thorndycraft V.R. 2005. Palaeoflood hydrology and its role in applied hydrological sciences. *Journal of Hydrology* 313: 3–15.

Bersezio, R., Giudici, M. and Mele, M. 2007. Combining sedimentological and geophysical data for high-resolution 3-D mapping of fluvial architectural elements in the Quaternary Po plain (Italy). *Sedimentary Geology* 202: 230–248.

Bettis, A.E. III. 1992. The soil morphologic properties and weathering zone characteristics as age indicators in Holocene alluvium in the upper Midwest. In: Holliday, V.T., ed., *Soils in Archeology, Landscape Evolution and Human Occupation*, Washington, DC: Smithsonian Institution, pp. 119–144.

Bierman, P.R. 1994. Using in situ produced cosmogenic isotopes to estimate rates of landscape evolution: a review from the geomorphic perspective. *Journal of Geophysical Research* 99: 13885–13896.

Birkeland, P.W. 1968. Mean velocities and boulder transport during Tahoe-age floods of the Truckee River, California–Nevada. *Geological Society of America Bulletin* 79: 137–142.

Birkeland, P.W. 1984. *Soils and Geomorphology*, New York: Oxford University Press, 372pp.

Bowman, S. 1990. *Radiocarbon Dating*, London: Trustees of the British Museum, 64pp.

Bradley, W.C. and Mears, A.I. 1980. Calculations of flows needed to transport coarse fraction of Boulder Creek alluvium at Boulder, Colorado. *Geological Society of America Bulletin, Part II* 91: 1057–1090.

Brakenridge, G.R. 1984. Alluvial stratigraphy and radiocarbon dating along the Duck River, *Tennessee. Implications regarding flood-plain origin. Geological Society of America Bulletin* 95: 9–25.

Bretz, J H. 1928. Bars of the channeled scabland. *Geological Society of America Bulletin* 39: 643–702.

Bridge, J.S. and Mackey, S.D. 1993. A revised alluvial stratigraphy model. In: Marzo, M. and Puigdefabregas, C., eds., Alluvial Sedimentation, *International Association of Sedimentologists Special Publication 17*, Oxford: Blackwell, pp. 319–336.

Brown, A.G. 1990. Holocene floodplain diachronism and inherited downstream variations in fluvial processes: a study of the river Perry, Shropshire, England. *Journal of Quaternary Science* 5: 39–51.

Brown, A.G. 1997. *Alluvial Geoarchaeology*, Cambridge: Cambridge University Press, 377pp.

Bull, W.B. 1990. Stream-terrace genesis: implications for soil development. In: Knuepher, P.L. and McFadden, L.D., eds., *Soils and Landscape Evolution: Proceedings of the 21st Annual Binghamton Symposium in Geomorphology*, Special Issue, *Geomorphology*, 3(3–4): 351–367.

Bull, W.B. 1991. *Geomorphic Responses to Climatic Change*, New York: Oxford University Press, 326pp.

Bullinger-Weber, G. and Gobat, J.-M. 2006. Identification of facies models in alluvial soil formation: the case of a Swiss alpine floodplain. *Geomorphology* 74: 181–195.

Chorley, R.J., Dunn, A.J. and Beckinsale, R.P. 1964. *The History of the Study of Landforms or the Development of Geomorphology*, vol. 1, London: Methuen, 678pp.

Church, M. 1978. Paleohydrological reconstructions from a Holocene valley fill. In: Miall, A.D., ed., *Fluvial Sedimentology*, Memoir 5, Calgary: Canadian Society of Petroleum Geologists, pp. 743–772.

Coates, D.R. 1984. Landforms and landscapes as measures of relative time. In: Mahaney, W.C., ed., *Quaternary Dating Methods*, New York: Elsevier, pp. 247–267.

Coleman, D.J. 1982. *An Examination of Bankfull Discharge Frequency in Relation to Floodplain Formation*, Baltimore, MD: Johns Hopkins University Press, 192pp.

Collier, M., Webb, R.H. and Schmidt, J.C. 1996. *Dams and Rivers*, US Geological Survey Circular 1126, Tucson, AZ: US Geological Survey, 94pp.

Costa, J.E. 1975. Effects of agriculture on erosion and sedimentation in the Piedmont Province, Maryland. *Geological Society of America Bulletin* 86: 1281–1286.

Costa, J.E. 1978. Holocene stratigraphy in flood frequency analysis. *Water Resources Research* 14: 626–632.

Costa, J.E. 1983. Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range. *Geological Society of America Bulletin* 94: 986–1004.

Daniels, R.B. and Hammer, R.D. 1992. *Soil Geomorphology*, New York: John Wiley & Sons, 236pp.

Dennis, I.A., Coulthard, T.J., Brewer, P. and Macklin, M.G. 2009. The role of floodplains in attenuating contaminated sediment fluxes in formerly mined drainage basins. *Earth Surface Processes and Landforms* 34:453–466.

Dorn, R.I. 1994. Surface exposure dating with rock varnish. In: Beck, C., ed., *Dating in Exposed and Surface Contexts*, Albuquerque, NM: University of New Mexico Press, pp. 77–113.

Dott, R.H. and Batten, R.L. 1976. *Evolution of the Earth*, New York: McGraw-Hill, 504pp.

Duller, G.A.T. 1996. Recent developments in luminescence dating of Quaternary sediments. *Progress in Physical Geography*, 20: 127–145.

Durham, R.W. and Joshi, S.R. 1984. Lead-210 dating of sediments from some northern Ontario lakes. In: Mahaney, W.C., ed., *Quaternary Dating Methods*, New York: Elsevier, pp. 75–85.

Dury, G.H. 1954. Contribution to a general theory of meandering valleys. *American Journal of Science* 252: 193–224.

Dury, G.H. 1965. *Theoretical Implications of Underlift Streams*, US Geological Survey Professional Paper 452-C, Washington, DC: US Government Printing Office, 43pp.

Dury, G.H. 1976. Discharge prediction, present and former, from channel dimensions. *Journal of Hydrology* 30: 219–245.

Dury, G.H. 1985. Attainable standards of accuracy in the retrodiction of palaeodischarge from channel dimensions. *Earth Surface Processes and Landforms* 10: 205–213.

Easterbrook, D.J., ed. 1988a. *Dating Quaternary Sediments*, Special Paper 227, Boulder, CO: Geological Society of America, 165pp.

Easterbrook, D.J. 1988b. Paleomagnetism of Quaternary deposits.In: Easterbrook, D.J., ed., Dating Quaternary Sediments, Special Paper 227, Boulder, CO: Geological Society of America, pp. 111–122.

Ely, L.L. and Baker, V.R. 1985. Reconstructing palaeoflood hydrology with slackwater deposits – Verde River, Arizona. *Physical Geography* 6: 103–126.

Ely, L.L., Enzel, Y., Baker, V.R. and Cayan, D.R. 1993. A 5000-year record of extreme floods and climate change in the southwestern United States. *Science* 262: 410–412.

Ely, L.L., Enzel, Y., Baker, V.R., Kale, V.S. and Mishra, S. 1996. Changes in the magnitude and frequency of late Holocene monsoon floods on the Narmada River, central India. *Geological Society of America Bulletin* 108: 1134–1148.

Ethridge, F.G. and Schumm, S.A. 1978. Reconstructing paleochannel morphologic and flow characteristics – methodology, limitations and assessment. In: Miall, A.D., ed., Fluvial Sedimentology, *Memoir 5*, Calgary: Canadian Society of Petroleum Geologists, pp. 703–721.

Fassnacht, H. 1998. *Frequency and Magnitude of Bedload Transport Downstream of the Pelton-Round Butte Dam Complex,* Lower Deschutes River, *Oregon*, MS Thesis, Oregon State University, 311pp.

Fassnacht.H., Klingeman, P.C. and Grant, G.E. 1998. Effects of a hydroelectric dam complex on bedload transport, lower Deschutes River, Oregon [abstract]. *EOS: Transactions of the American Geophysical Union* 79: F374.

Ferring, C.R. 1992. Alluvial pedology and geoarcheological research. In: Holliday, V.T., ed., *Soils in Archeology, Landscape Evolution and Human Occupation*, Washington, DC: Smithsonian Institution, pp. 1–39.

Fredlund, G.G. 1996. Late Quaternary geomorphic history of lower Highland Creek, Wind Cave National Park, South Dakota. *Physical Geography* 17: 446–464.

Gaffney, C. 2008. Detecting trends in the prediction of the buried past: a review of geophysical techniques in archaeology. *Archaeometry* 50: 313–336.

Gilbert, G.K. 1914. The Transportation of Debris by Running Water, *US Geological Survey Professional Paper 86*, Washington, DC: US Government Printing Office, 263pp.

Grant, G.E. 1997. Critical flow constrains flow hydraulics in mobile-bed streams – a new hypothesis. *Water Resources Research* 33: 349–358.

Gregory, J.J. and Maizels, J.K. 1991. Morphology and sediments: typological characteristics of fluvial forms and deposits. In: Starkel, L., Gregory, K.J. and Thornes, J.B., eds., *Temperate Palaeohydrology*, Chichester: John Wiley & Sons, pp. 31–59.

Gruen, R. 1994, Quaternary geochronology. *Quaternary Science Reviews* 13(2): 95–181.

Hallet, B. and Putkonen, J. 1994. Surface dating of dynamics landforms; young boulders on aging moraines. *Science* 265: 937–940.

Harden, T.M., Macklin, M.G. and Baker, V.R. 2010. Holocene flood histories in south-western USA. *Earth Surface Processes and Landforms* 35: 707–716.

Harden, T.M., O'Connor, J.E., Driscoll, D.G. and Stamm, J.F. 2011. *Flood-frequency Analyses from Paleoflood Investigations for Spring, Rapid, Boxelder and Elk Creeks, Black Hills, Western South Dakota*, USGS Scientific Investigations Report 2011-5131, Reston, VA: US Geological Survey, 136pp.

Harvey, J.E. and Pederson, J.L. 2011. Reconciling arroyo cycle and paleoflood approaches to late Holocene alluvial records in dryland streams. *Quaternary Science Reviews* 30:55–866.

Harvey, J.E., Pederson, J.L. and Rittenour, T.M. 2011. Exploring relations between arroyo cycles and canyon paleoflood records in Buckskin Wash, Utah: reconciling scientific paradigms. *Geological Society of America Bulletin* 123: 2266–2276.

Haynes, C.V. 1985. *Mastodon-bearing Springs and Late Quaternary Geochronology of the Lower Pomme de Terre Valley, Missouri*, Special Paper 204, Boulder, CO: Geological Society of America, 35pp.

Hidy, A.J., Gosse, J.C., Pederson, J.L., Mattern, J.P. and Finkel, R.C. 2010. A geologically constrained Monte Carlo approach to modeling exposure ages from profiles of cosmogenic nuclides: an example from Lees Ferry, Arizona. *Geochemistry, Geophysics, Geosystems* 11: Q0AA10.

Holliday, V.T. 1990. Pedology in archaeology. In: Lasca, N.P. and Donahue, J., eds., *Archaeological Geology of North America: Decade of North American Geology*, Centennial Special Volume 4, Boulder, CO: Geological Society of America, pp. 525–540.

Hosman, K.L., Ely, L.L. and O'Connor, J.E. 2003. Holocene paleoflood hydrology of the Lower Deschutes River, Oregon. In: O'Connor, J.E. and Grant, G.E., eds., A Peculiar River: Geology, Geomorphology and Hydrology of the Deschutes River, Oregon, *Water Science and Application Series*, vol. 7, Washington, DC: American Geophysical Union, pp. 121–146.

House P.K., Webb, R.H., Baker, V.R., Levish D.R., eds. 2002. *Ancient Floods, Modern Hazards: Principles and Applications of Paleoflood Hydrology*. Water Science and Application Series, vol. 5, Washington, DC: American Geophysical Union, 385pp.

Hughes, P.D. 2010. Geomorphology and Quaternary stratigraphy: the roles of morpho-, litho- and allostratigraphy. *Geomorphology* 123: 189–199.

Hupp, C.R. 1988. Plant ecological aspects of flood geomorphology and paleoflood history. In: Baker, V.R., Kochel, R.C. and Patton, P.C., eds., *Flood Geomorphology*, New York: John Wiley & Sons, pp. 335–356.

Hutton, J. 1788, Theory of the Earth. *Royal Society of Edinburgh, Transactions* 1: 209–304.

International Subcommission on Stratigraphic Classification. 1994. *A Guide to Stratigraphic Classification, Terminology and Procedure*, Trondheim, Norway and Boulder, CO: International Union of Geological Sciences and Geological Society of America, 214 pp.

Iwai, S. and Ishiguro, M. 1970. *Applied Statistics for Hydrology*, Tokyo: Morikita Shuppan, 369 pp. (in Japanese).

Jacobson, R.B. 2004. Watershed sustainability: downstream effects of timber harvest in the Ozarks of Missouri. In: Flader, S.J., ed., *Toward Sustainability for Missouri Forests*, General Technical Publication NC-239, St Paul, MN: USDA Forest Service North Central Research Station, pp. 106–128.

Jacobson, R.B. and Coleman, D.J. 1986. Stratigraphy and recent evolution of Maryland Piedmont floodplains. *American Journal of Science* 286: 617–637.

Jacobson, R.B. and Gran, K.B. 1999. Gravel routing from widespread, low-intensity landscape disturbance, Current River Basin, Missouri. *Earth Surface Processes and Landforms* 24: 897–917.

Jacobson, R.B. and Primm, A.T. 1997. *Historical Land-use Changes and Potential Effects on Stream Disturbance in the Ozark Plateaus, Missouri*, US Geological Survey Water Supply Paper 2484, Washington, DC: US Government Printing Office, 85pp.

Jacobson, R.B. and Pugh, A.L. 1992. Effects of land use and climate shift on channel disturbance, Ozark Plateaus, USA. In: *Proceedings of the Workshop on the Effects of Global Climate Change on Hydrology and Water Resources at the Catchment Scale*, Japan–US Committee on Hydrology, Water Resources and Global Climate Change, pp. 432–444.

Jacobson, R.B. and Pugh, A.L. 1997. Riparian vegetation and the spatial pattern of stream-channel instability, Little Piney Creek, Missouri. US Geological Survey Water Supply Paper 2494, Washington, DC: US Government Printing Office, 33pp.

Jenny, H. 1941. *Factors of Soil Formation*, New York: McGraw-Hill, 281pp.

Johnson, M.D., Jennings, C. and Hobbs, H. 2009. Lithostratigraphy for Quaternary glacial deposits: "If it ain't broke, don't fix it!", *GSA Today* 9: e15–e16.

Jones, A.P., Shimazu, H., Oguchi, T., Okuno, M. and Tokutake, M. 2001. Late Holocene slackwater deposits on the Nakagawa River, Tochigi Prefecture, Japan. *Geomorphology* 39: 39–51.

Katsube, K. and Oguchi, T. 1999. Altitudinal changes in slope angle and profile curvature in the Japan Alps: a hypothesis regarding a characteristic slope angle. *Geographical Review of Japan* 72B: 63–72.

Kehew, A.E. and Teller, J.T. 1994. Glacial-lake spillway incision and deposition of a coarse-grained fan near Watrous, Saskatchewan. *Canadian Journal of Earth Sciences* 31: 544–553.

Kemp, P.A. 1985. *Soil Micromorphology and the Quaternary*, Amsterdam: Elsevier, 79pp.

Kite, J.S. and Linton, R.C. 1991. Depositional aspects of the November 1985 flood on Cheat River and Black Fork, West Virginia. In: Jacobson, R.B., ed., *Geomorphic Studies of the Storm and Flood of November 3–5 1985, in the Upper Potomac and Cheat River Basins in West Virginia and Virginia*, US Geological Survey Bulletin 1981, p. D21–D24.

Kleinhans, M.G. and van den Berg, J.H. 2011. River channel and bar patterns explained and predicted by an empirical and physics-based method. *Earth Surface Processes and Landforms* 36:721–738.

Knox, J.C. 1988. Climatic influence on upper Mississippi valley floods. In: Baker, V.R., Kochel, R.C. and Patton, P.C., eds., *Flood Geomorphology*, New York: John Wiley & Sons, pp. 279–300.

Knox, J.C. 1993. Large increases in flood magnitude in response to modest changes in climate. *Nature* 361: 420–432.

Knox, J.C. 2006. Floodplain sedimentation in the Upper Mississippi Valley: natural versus human accelerated. *Geomorphology* 79: 286–310.

Knuepfer, P.L. 1994. Use of rock weathering rinds in dating geomorphic surfaces. In: Beck, C., ed., *Dating in Exposed and Surface Contexts*, Albuquerque, NM: University of New Mexico Press, pp. 15–28.

Kochel, R.C. and Baker, V.R. 1982. Paleoflood hydrology. *Science* 215: 353–361.

Kochel, R.C. and Baker, V.R. 1988. Paleoflood analysis using slackwater deposits. In: Baker, V.R., Kochel, R.C. and Patton, P.C., eds., *Flood Geomorphology*, New York: John Wiley & Sons, pp. 357–376.

Kochel, R.C., Baker, V.R. and Patton, P.C. 1982. Paleohydrology of southwestern Texas. *Water Resources Research* 18: 1165–1183.

Kohler, S. 1984. *Two Ozark Rivers: the Current and the Jacks Fork*, Columbia, MO: University of Missouri Press, 130pp.

Komar, P.D. 1970. The competence of turbidity current flow. *Geological Society of America Bulletin* 81: 1555–1562.

Komar, P.D. 1989. Flow-competence evaluations of the hydraulic parameters of floods – an assessment of the technique. In: Beven, K. and Carling, P., eds., *Floods – Hydrological,* Sedimentological and Geomorphological Implications, Chichester: John Wiley & Sons, pp. 107–134.

Komar, P.D. 1996. Entrainment of sediments from deposits of mixed grain sizes and densities. In: Carling, P.A. and Dawson, M.R., eds., *Advance in Fluvial Dynamics and Stratigraphy*, Chichester: John Wiley & Sons, pp. 107–134.

Komar, P.D. and Carling, P.A. 1991. Grain sorting in gravel-bed streams and the choice of particle sizes for flow-competence evaluations. *Sedimentology* 38: 489–502.

Kraus, M.J. and Brown, T.M. 1988. Pedofacies analysis: a new approach to reconstructing ancient fluvial sequences. In: Reinhardt, J. and Sigleo, W.R., eds., *Paleosols and Weathering Through Geologic Time: Principles and Applications*, Special Paper 216, Boulder, CO: Geological Society of America, pp. 143–152.

Kurz, M.D. and Brook, E.J. 1994. Surface exposure dating with cosmogenic nuclides. In: Beck, C., ed., *Dating in Exposed and Surface Contexts*, Albuquerque, NM: University of New Mexico Press, pp. 139–159.

Lamb, M.P. and Fonstad, M.A. 2010. Rapid formation of a modern bedrock canyon by a single flood event. *Nature Geoscience* 3: 477–481.

Lasca, N.P. and Donahue, J. 1990. Archaeological Geology of North America: Decade of North American Geology, *Centennial Special Volume 4*, Boulder, CO: Geological Society of America, 633pp.

Laustela, M., Egli, M., Frauenfelder, R., Kaab, A., Maisch, M. and Haeberli, W. 2004. Weathering rind measurements and relative age dating of rock glacier surfaces in crystalline regions of the Eastern Swiss Alps. In: Phillips, M., Springman, S.M. and Arenson, L.U., eds., *Permafrost: Proceeding of the 8th International Conference on Permafrost, 2003*, Boca Raton, FL: Taylor and Francis, pp. 627–632.

Leopold, L.B. and Wolman, M.G. 1957. *River Channel Patterns: Braided, Meandering and Straight*, US Geological Survey Professional Paper 282-B, Washington, DC: US Government Printing Office, 39pp.

Leopold, L.B., Wolman, M.G. and Miller, J.P. 1964. *Fluvial Processes in Geomorphology*, San Francisco, CA: W.H. Freeman, 522pp.

Levish, D.R. and Ostenna, D.A. 1996. Applied paleoflood hydrology in north-central Oregon (Bureau of Reclamation Seismotectonic Report 96-7), Portland, OR: *Guidebook for Field Trip 2, Cordilleran Section Meeting of the Geological Society of America*, unpaginated.

Lewin, J., Macklin, M.G. and Johnstone, E. 2005. Interpreting alluvial archives: sedimentological factors in the British Holocene fluvial record. *Quaternary Science Reviews* 24: 1873–1889.

Lord, M.L. and Kehew, A.E. 1987. Sedimentology and paleohydrology of glacial-lake outburst deposits in southeastern Saskatchewan and northwestern North Dakota. *Geological Society of America Bulletin* 99: 663–673.

Love, K. 1990. Paradise lost. *Missouri Conservationist* 51: 31–35.

Lund, S.P. 1996. A comparison of Holocene paleomagnetic secular variation records from North America. *Journal of Geophysical Research, B, Solid Earth and Planets* 101: 8007–8024.

Macklin, M.G. and Rumsby, B.T. 2007. Changing climate and extreme floods in the British uplands. *Transactions of the Institute of British Geographers* 32: 168–186.

Macklin, M.G., Jones, A.F. and Lewin, J. 2010. River response to rapid Holocene environmental change: evidence and explanation in British catchments. *Quaternary Science Reviews* 29: 1555–1576.

Magilligan, F.J. 1985. Historical floodplain sedimentation in the Galena River Basin, Wisconsin and Illinois. *Annals of the Association of American Geographers* 75: 583–594.

Mahaney, W.C. 1984. *Quaternary Dating Methods*, New York: Elsevier, 431pp.

Maizels, J.K. 1983. Palaeovelocity and palaeodischarge determination for coarse gravel deposits. Gregory, K.J., ed., *Background to Palaeohydrology*, Chichester: John Wiley & Sons, pp. 101–139.

Malde, H.E. 1968. *The Catastrophic Late Pleistocene Bonneville Flood in the Snake River Plain, Idaho*, US Geological Survey Professional Paper 596, Washington, DC: US Government Printing Office, 52pp.

Malmon, D.V., Reneau, S.L., Dunne, T., Katzman, D. and Drakos, P.G. 2005. Influence of sediment storage on downstream delivery of contaminated sediment. *Water Resources Research* 41: W05008.

Markewich, H.W. and Pavich, M.J. 1991. Soil chronosequence studies in temperate to subtropical, low-latitude, low-relief terrain with data from the Eastern United States. In: Pavich, M.J., ed., *Weathering and Soils*, Special Issue, *Geoderma* 51(1–4): 213–239.

Marron, D.C. 1992. Floodplain storage of mine tailings in the Belle Fourche River system: a sediment budget approach. *Earth Surface Processes and Landforms* 17: 675–685.

Martin, C.W. 1992. The response of fluvial systems to climate change; an example from the central Great Plains. *Physical Geography* 13: 101–114.

Marzo, M. and Puigdefabregas, C., eds. 1993. Alluvial Sedimentation, *International Association of Sedimentologists Special Publication 17*, Oxford: Blackwell, 586pp.

Matsumoto, J. 1993. Global distribution of daily maximum precipitation. *Bulletin of the Department of Geography, University of Tokyo* 25: 43–48.

Matthews, J.A. 1994. Lichenometric dating: a review with particular reference to 'Little Ice Age' moraines in southern Norway. In: Beck, C., ed., *Dating in Exposed and Surface Contexts*, Albuquerque, NM: University of New Mexico Press, pp. 185–212.

McClure, E.M. 1998. *Spatial and Temporal Trends in Bed Material and Channel Morphology Below a Hydroelectric Dam Complex,* Deschutes River,Oregon, MS Thesis, Oregon State University, 85pp.

McClure, E.M., Grant, G.E. and Jones, J.A. 1997. Longitudinal patterns of bed material size following impoundment of the lower Deschutes River, Oregon [abstract]. *Geological Society of America Abstracts With Programs* 29: 314.

McKenney, R. 1997. *Formation and Maintenance of Hydraulic Habitat Units in the Streams of the Ozark Plateaus, Missouri and Arkansas*, PhD Dissertation, State College, PA: Pennsylvania State University, 347pp.

McKenney, R., Jacobson, R.B. and Wertheimer, R.C. 1995. Woody vegetation and channel morphogenesis in low-gradient, gravel-bed streams in the Ozarks Region, Missouri and Arkansas. *Geomorphology* 13: 175–198.

Miall, A.D. 2010. *The Geology of Fluvial Deposits: Sedimentary Facies,* Basin Analysis and Petroleum Geology, New York: Springer, 598pp.

Miller, H. 1970. Methods and results of river terracing. In: Dury, G.H., ed., *Rivers and River Terraces*, New York: Praeger, pp. 19–35.

Nanson, G.C., Chen, X.Y. and Price, D.M. 1995. Aeolian and fluvial evidence of changing climate and wind patterns during the past 100 ka in the western Simpson Desert, Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 113: 87–102.

Nikiforoff, C.C. 1949. Weathering and soil evolution. *Soil Science* 67: 219–223.

Nolan, K.M., Kelsey, H.M. and Marron, D.C. 1995. *Geomorphic Processes and Aquatic Habitat in the Redwood Creek Basin, Northwestern California*, US Geological Survey Professional Paper 1454, Washington, DC: US Government Printing Office, pp. A1–A6.

North American Commission on Stratigraphic Nomenclature. 1983. North American Stratigraphic Code. *American Association of Petroleum Geologists Bulletin* 67: 841–875.

O'Connor, J.E. 1993. *Hydrology, Hydraulics and Geomorphology of the Bonneville Flood*, Special Paper 274, Boulder, CO: Geological Society of America, 83pp.

O'Connor, J.E. and Grant, G.E., eds. 2003. *A Peculiar River: Geology, Geomorphology and Hydrology of the Deschutes River, Oregon*, Water Science and Application Series, vol. 7, Washington, DC: American Geophysical Union, 219pp.

O'Connor, J.E., Webb, R.H. and Baker, V.R. 1986. Paleohydrology of pool-and-riffle pattern development: Boulder Creek, Utah. *Geological Society of America Bulletin* 97: 410–420.

O'Connor, J.E., Ely, L.L., Wohl, E.E., Stenvens, L.E., Melis, T.S., Kale, V.S. and Baker, V.R. 1994. A 4500-year record of large floods on the Colorado River in the Grand Canyon, Arizona. *Journal of Geology* 102: 1–9.

O'Connor, J.E., Grant, G.E., Curran, J.H., Fassnacht.H. and Brink, M. 1998. Geomorphic processes within the lower Deschutes River, Oregon [abstract]. *EOS: Transactions of the American Geophysical Union* 79: F374.

Oguchi, T. 1996. Hillslope failure and sediment yield in Japanese regions with different storm intensity. *Bulletin of the Department of Geography, University of Tokyo* 28: 45–54.

Oguchi, T. 1997. Late Quaternary sediment budget in alluvial-fan–source-basin systems in Japan. *Journal of Quaternary Science* 12: 381–390.

Owens, P.N., Walling, D.E. and Leeks, G.J.L. 1999. Use of floodplain sediment cores to investigate recent historical changes in overbank sedimentation rates and sediment sources in the catchment of the River Ouse, Yorkshire. *Catena* 36: 21–47.

Parker, G. 1976. On the cause and characteristic scales of meandering and braiding in rivers. *Journal of Fluid Mechanics* 76: 457–480.

Partridge, J. and Baker, V.R. 1987. Paleoflood hydrology of the Salt River, Arizona. *Earth Surface Processes and Landforms* 12: 109–125.

Patton, P.C. and Dibble, D.S. 1982. Archeologic and geomorphic evidence for paleohydrologic record of the Pecos River in west Texas. *American Journal of Science* 282: 97–121.

Patton, P.C., Baker, V.C. and Kochel, R.C. 1979. Slack-water deposits; a geomorphic technique for the interpretation of fluvial paleohydrology. In: Rhodes, D.D. and Williams, G.P., eds., *Adjustments of the Fluvial System, a Proceedings Volume of the Annual Geomorphology Symposia Series, Tenth Annual Geomorphology Symposium, held at Binghampton, New York, September 21–22, 1979*, Dubuque, IA: Kendall-Hunt, pp. 225–253.

Pazzaglia, F.J., Gardner, T.W. and Merritts, D.J. 1998. Bedrock fluvial incision and longitudinal profile development over geologic time scales determined by fluvial terraces. In: Tinkler, K.J. and Wohl, E.E., eds., *Rivers Over Rock: Fluvial Process in Bedrock Channels*, Geophysical Monograph 107, Washington, DC: American Geophysical Union, pp. 207–235.

Peterson, J.T. 1996. *The Evaluation of a Hydraulic Unit-based Habitat System*, PhD Dissertation, Columbia, MO: University of Missouri, 397pp.

Pettijohn, F.J. 1975. *Sedimentary Rocks*, New York: Harper and Row, 628pp.

Pinter, N., Keller, E.A. and West, R.B. 1994. Relative dating of terraces of the Owens River, northern Owens Valley, California and correlation with moraines of the Sierra Nevada. *Quaternary Research* 42: 266–276.

Rasanen, M.E., Auri, J.M., Huitti, J.V., Klap, A.K. and Virtsasalo, J.J. 2009. A shift from lithostratigraphic to allostratigraphic classification of Quaternary glacial deposits. *GSA Today* 19: 4–11

Reading, H.G. 1978. *Sedimentary Environments and Facies*, New York: Elsevier, 557pp.

Reynolds, J.M. 2011. *An Introduction to Applied and Environmental Geophysics*, 2nd edn., Chichester: John Wiley & Sons, 712pp.

Rice, S. and Church, M. 1998. Grain size along two gravel-bed rivers; statistical variation, spatial pattern and sedimentary links. *Earth Surface Processes and Landforms* 23: 345–363.

Rittenour, T.M. 2008. Luminescence dating of fluvial deposits: applications to geomorphic, palaeoseismic and archaeological research. *Boreas* 37: 613–635.

Robinson, D.A., Binley, A., Crook, N., Day-Lewis, F.D., Ferré, T.P.A., Grauch, V.J.S., Knight, R., Knoll, M., Lakshmi, V., Miller, R., Nyquist, J., Pellerin, L., Singha, K. and Slater, L. 2008. Advancing process-based watershed hydrological research using near-surface geophysics: a vision for, and review of, electrical and magnetic geophysical methods. *Hydrological Processes* 22: 3604–3635.

Rotnicki, K. 1983. Modelling past discharges of meandering rivers. In: Gregory, K.J., ed., *Background to Palaeohydrology*, Chichester: John Wiley & Sons, pp. 321–354.

Royall, P.D., Delcourt, P.A. and Delcourt, H.R. 1991. Late Quaternary paleoecology and paleoenvironments of the central Mississippi alluvial valley. *Geological Society of America Bulletin* 103: 157–170.

Ruhe, R.V. 1975. *Geomorphology –Geomorphic Processes and Surficial Geology*, Boston, MA: Houghton Mifflin, 246pp.

Saito, K. 1988. *Alluvial Fans in Japan*. Tokyo: Kokon-Shoin, 280pp. (in Japanese).

Sarna-Wojcicki, A.M., Bowman, H.R., Meyer, C.E., Russell, P.C., Woodward, M.J., McCoy, G., Rowe, J.J. Jr, Baedecker, P.A., Asaro, F. and Michael, H. 1984. *Chemical Analyses, Correlations and Ages of Upper Pliocene and Pleistocene Ash Layers of East-Central and Southern California*. US Geological Survey Professional Paper 1293, Washington, DC: US Government Printing Office, 40pp.

Sarna-Wojcicki, A.M., Lajoie, K.R., Meyer, C.E., Adam, D.P. and Rieck, H.J. 1991. Tephrochronologic correlation of upper Neogene sediments along the Pacific margin, conterminous United States. In: Morrison, R.B., ed., Quaternary Nonglacial Geology; Conterminous U.S.: Decade of North American Geology, vol. K-2, Boulder, CO: Geological Society of America, pp. 117–140.

Schmidt, S., Hetzel, R., Kuhlmann, J., Mingorance, F. and Ramos, V.A. 2011. A note of caution on the use of boulders for exposure dating of depositional surfaces. *Earth and Planetary Science Letters* 302: 60–70.

Schneider, J.S. and Bierman, P.R. 1997. Surface dating using rock varnish. In: Taylor, R.E. and Aitken, M.J., eds., *Chronometric Dating in Archaeology*, New York: Plenum Press, pp. 357–388.

Schumm, S.A. 1967. Meander wavelength of alluvial rivers. *Science* 157: 1549–1550.

Schumm, S.A. 1968. *River Adjustment to Altered Hydrologic Regimen – Murrumbidgee River and Paleochannels, Australia*. US Geological Survey Professional Paper 598, Washington, DC: US Government Printing Office, 65pp.

Schumm, S.A. and Khan, H.R. 1972. Experimental study of channel patterns. *Geological Society of America Bulletin* 83: 1755–1770.

Schumm, S.A. and Parker, R.S. 1973. Implications of complex response of drainage systems for Quaternary alluvial stratigraphy. *Nature* 243: 99–100.

Sharma, P.V. 1997. *Environmental and Engineering Geophysics*. Cambridge: Cambridge University Press, 475pp.

Sheldon, N.D. and Tabor, N.J. 2009. Quantitative paleoenvironmental and paleoclimatic reconstruction using paleosols. *Earth Science Reviews* 95: 1–52.

Smith, N.D. and Rogers, J., eds. 2010. *Fluvial Sedimentology VI*. Special Publication 28 of the International Association of Sedimentologists: Hoboken, NJ: Wiley-Blackwell, 488pp.

Sternberg, R.S. and McGuire, R.H. 1990. Secular variation in the American southwest. In: Eighmy, J.L. and Sternberg, R.S., eds., *Archaeomagnetic Dating*, Tucson, AZ: University of Arizona Press, pp. 199–225.

Stuiver, M. 1982. A high-precision calibration of the AD radiocarbon time scale. *Radiocarbon* 24: 1–26.

Stuiver, M. and Kra, R.S., eds. 1986. Calibration Issue. *Radiocarbon* 28: 805–1030.

Stupavsky, M. and Gravenor, C.P. 1984. Paleomagnetic dating of Quaternary sediments: a review. In: Mahaney, W.C., ed., *Quaternary Dating Methods*, New York: Elsevier, pp. 123–140.

Sugai, T. 1993. River terrace development by concurrent fluvial processes and climatic changes. *Geomorphology* 6: 243–252.

Surovell, T.A., Finley, J.B., Smith, G.M., Brantingham, P.J. and Kelly, R. 2009. Correcting temporal frequency distributions for taphonomic bias. *Journal of Archaeological Science* 36: 1715–1724.

Suzuki, H. 1971. Climatic zones of the Wurm Glacial Age. *Bulletin of the Department of Geography, University of Tokyo* 3: 35–46.

Taylor, M.P. and Lewin, J. 1996. River behavior and Holocene alluviation: the River Severn at Welshpool, mid-Wales, U.K. *Earth Surfaces Processes and Landforms* 21: 77–91.

Taylor, R.E., Long, A. and Kra, R.S. 1992. *Radiocarbon After Four Decades – An Interdisciplinary Perspective*, New York: Springer, 596pp.

Tokyo Bureau of International Trade and Industry. 1984. *Investigation Report Toward Appropriate Ground Water Usage in the Minimiazumi District, Nagano Prefecture*, Tokyo: Bureau of International Trade and Industry, 120pp. (in Japanese).

Trimble, S.W. 1974. *Man-induced Soil Erosion on the Southern Piedmont 1700–1970*, Ankeny, IA, Soil Conservation Society of America, 180pp.

Trimble, S.W. and Lund, S.W. 1982. *Soil Conservation and Reduction of Erosion and Sedimentation in the Coon Creek Basin, Wisconsin*, US Geological Survey Professional Paper 1234, Washington, DC: US Government Printing Office, 35pp.

US Army Corps of Engineers. 1998. *Geophysical Exploration for Engineering and Environmental Investigations*, Reston, VA: American Society of Civil Engineers Press, 204pp.

Verosub, K.L. 1988. Geomagnetic secular variation and the dating of Quaternary sediments. In: Easterbrook, D.J., ed., *Dating Quaternary Sediments*, Special Paper 227, Boulder, CO: Geological Society of America, pp. 123–138.

Vincent, K.R., Bull, W.B. and Chadwick, O.A. 1994. Construction of a soil chronosequence using the thickness of pedogenic carbonate coatings. *Journal of Geological Education* 42: 316–324.

Walker, R.G. and James, N.P. 1992. *Facies Models – Response to Sea Level Change*, St. *John's*, Newfoundland: Geological Society of Canada, 454pp.

Walter, R.C. and Merritts, D.J. 2008. Natural streams and the legacy of water-powered mills. *Science* 319: 299–304.

Waythomas, C.F., Walder, J.S., McGimsey, R.G. and Neal, C.A. 1996. A catastrophic flood caused by drainage of a caldera lake at Aniakchak Volcano, Alaska, and implications for volcanic hazards assessment. *Geological Society of America Bulletin* 108: 861–871.

Webb, R.H. 1985. *Late Holocene Flooding on the Escalante River,* South-central Utah, PhD Dissertation, Tucson, AZ: University of Arizona, 204pp.

Webb, R.H. and Jarrett, R.D., 2002. One-dimensional estimation techniques for discharges of paleofloods and historical floods. In: House, P.K., Webb, R.H., Baker, V.R. and Levish, D.R., eds., *Ancient Floods, Modern Hazards: Principles and Applications of Paleoflood Hydrology*, Water Science and Application Series, vol. 5, Washington, DC, American Geophysical Union, pp. 111–125

Webb, R.H., O'Connor, J.E. and Baker, V.R. 1988. Paleohydrologic reconstruction of flood frequency on the Escalante River, south-central Utah. In: Baker, V.R., Kochel, R.C. and Patton, P.C., eds., *Flood Geomorphology*, New York: John Wiley & Sons, pp. 403–418.

Weissel, J.K. and Seidel M.A. 1998. Inland propagation of erosional escarpments and river profile evolution across the southeast Australia passive continental margin. In: Tinkler, K.J. and Wohl, E.E., eds., *Rivers Over Rock: Fluvial Process in Bedrock Channels*, Washington, DC: American Geophysical Union, pp. 189–296.

Wells, L.W. 1990. Holocene history of the El Niño phenomenon as recorded in flood sediments of northern coastal Peru. *Geology* 18: 1134–1137.

Wilcock, P.R. 1992. Flow competence: a criticism of a classic concept. *Earth Surface Processes and Landforms* 7: 289–298.

Williams, G.P. 1978. Bankfull discharge of rivers. *Water Resources Research* 14: 1141–1154.

Williams, G.P. 1983. Paleohydrological methods and some examples from Swedish fluvial environments. I. Cobble and boulder deposits. *Geografiska Annaler* 65A: 227–243.

Williams, G.P. 1984a. Paleohydrologic equations for rivers. In: Costa, J.E. and Fleisher, P.J., eds., *Developments and Applications of Geomorphology*, New York: Springer, pp. 343–367.

Williams, G.P. 1984b. Paleohydrological methods and some examples from Swedish fluvial environments. II. River meanders. *Geografiska Annaler* 66A: 89–102.

Williams, G.P. 1988. Paleofluvial estimates from dimensions of former channels and meanders. In: Baker, V.R., Kochel, R.C. and Patton, P.C., eds., *Flood Geomorphology*, New York: John Wiley & Sons, pp. 321–334.

Wohl, E.E. 1988. *Northern Australian Paleofloods as Paleoclimatic Indicators*, PhD Dissertation, Tucson, AZ: University of Arizona, 285pp.

Wolcott, J. and Church, M. 1991. Strategies for sampling spatially heterogeneous phenomena; the example of river gravels. *Journal of Sedimentary Petrology* 61: 534–543.

Wolman, M.G. 1967. A cycle of sedimentation and erosion in urban river channels. *Geografiska Annaler* 49A: 385–395.

CHAPTER 3

Archaeology and human artefacts

Anthony G. Brown[1], François Petit[2] and L. Allen James[3]
[1]University of Southampton, Southampton, UK
[2]Institut de Géographie, Université de Liège – Sart Tilman, Liège, Belgium
[3]University of South Carolina, Columbia, Columbia, SC, USA

3.1 Introduction

Geomorphology and archaeology have strong historical and methodological links. Indeed, the origins of both geomorphology and archaeology lie in 18th and 19th century geology. The sub-discipline of geoarchaeology, defined as the use of geological and geomorphological methods in archaeology, has a long history (Zeuner 1945), even if the term is relatively new and of North American origin (Brown 1997). Geoarchaeology and its variants, archaeogeology (Renfrew 1976) and archaeological geology (*sensu* Herz and Garrison 1998), all seek to answer archaeological problems using techniques from the Earth Sciences (Waters 1992). The subject of this chapter is subtly different, as it is the use of archaeological evidence to answer questions concerning Earth surface processes and history, i.e. geomorphology. Although the most obvious way in which archaeology can be a tool in geomorphology is by dating sedimentation or erosion and thereby establishing rates of flux, there are many more applications including the identification and reconstruction of forcing factors on the Earth system (e.g. climate) and the history of human influences on Earth surface processes.

Archaeology can date erosion or deposition at the 10^3–10^4 year time-scale, and occasionally the temporal phasing of sites can be converted into a spatial phasing of erosional or depositional segments of the landscape, providing rates of erosion or deposition through site formation and destruction processes (Schiffer 1987). Examples include the use of tells or house mounds for estimating erosion rates (Kirkby and Kirkby 1976), the use of site distribution for erosional surveys (Thornes and Gilman 1983) or the use of artefacts and sites in the studies of river channel changes (Brown 2008). A second value of archaeological data is their potential to provide information concerning processes and environmental change. This approach has a venerable history in North America and particularly the American South West (Antevs 1935, 1955; Holliday 1992; Waters 1992). Environmental change, and particularly denudation history, in the Mediterranean has also been approached using archaeological data (Vita-Finzi 1969; Brown 2000) whereas only more recently have archaeological tools been extensively used in northwest Europe and the rest of the world. Environmental histories often reveal the role of humans in modifying of the physical environment, to the point where it is now believed that human agency may be the dominant geomorphological factor, in mass-flux terms, on the surface of the Earth in what has been termed 'The Anthropocene' period (Zalasiewicz *et al.* 2011). But for a geomorphological critique of this idea see Brown *et al.* (2016).

3.2 General considerations in using archaeological evidence in geomorphology

Artefacts can give geomorphologists a datum point and sometimes a date that is either imprecise (e.g. Mesolithic) or remarkably precise – sometimes even a calendar date (e.g. from inscriptions and coins). However, the values of the datum and date are dependent upon the origin of the artefact – if in situ it may record a landsurface, or if transported a depositional event. Clearly, the transport history of an artefact is a function of its origin and mass; small pottery shards are easily transported by rivers whereas blockwork from stone bridge piers rarely travels far. It must also be remembered that the discovery and use of such data are often a function of system behaviour such as the retreat of a river bank section or gully incision. As archaeological evidence is commonly found in floodplain sediments, tools used for analysing floodplain sedimentation (Chapter 2) are applicable. Organic artefacts may also be found, either preserved through desiccation (e.g. from human bodies to seeds) or waterlogging preventing aerobic decay. Examples include wooden artefacts such as bowls, tools and figurines (Coles and Coles 1989) and basketwork as associated with Mesolithic (Middle Stone Age) fishing on the Seine, France (Mordant and Mordant 1992) and Neolithic (New Stone Age) fishing on the River Trent, England (Brown 2009a). Environmental materials such as timbers, dated by ^{14}C and dendrochronology methods, have been used in the construction of a sedimentary model of river confluence evolution also on the River Trent (Brown *et al.* 2013). Animal and human bones may also be preserved as in the Bronze Age log jam in the River Trent, UK (Howard *et al.*

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

2000) There is clearly a preservation bias here and this greatly reduces the occurrence of such evidence in seasonally variable and tropical climates.

The most common and most valuable source of archaeological data is the exposure or section face. Visual searching can in the case of coins or other metal objects be augmented by the use of a metal detector. Finds should be located precisely on field drawings and depth logged. Artefacts on the ground surface, residual in archaeological terminology, have little dating value. Spot dating of pottery can be difficult or impossible and it should be remembered that much archaeological dating relies on associations between artefacts and is therefore not completely reliable.

3.3 Archaeological tools

A summary of the most common archaeological tools is given in Table 3.1 along with their potential and disadvantages. A more detailed discussion including selected examples is given below.

Hearths and lithics

In many sedimentary sequences in both the Old and New World hearths, charcoal from hearths and stone tools can be used as stratigraphic markers. Although charcoal is chemically ideal for conventional radiocarbon dating the most common problem is reworking so the charcoal must be in situ, otherwise it can only yield a maximum age for deposition. An example of a study that uses charcoal dating of hearths is the work of Baker *et al.* (1985) on slack-water deposits in Northern Territories, Australia. Indeed, in most studies of slack-water deposits, charcoal is the major material dated. It has also been extensively used in the determination of alluvial chronologies of arroyos in the southwestern United States (Waters and Nordt 1995). Flints can be dated using electron spin resonance, but it is rarely of sufficient precision for geomorphological studies. It is also theoretically possible to date artefacts using cosmogenic radionuclides (CRNs, e.g. ^{10}Be and ^{26}Al) (Brown 2011a); however, this offers no advantages and several disadvantages over CRN dating of other clasts within the sediment body.

Pottery and small artefacts

Pottery is often the most common artefact in sediments after ceramic civilizations appear. If large enough and of a diagnostic type it can be dated, although there are problems with on-site so-called 'spot dating' as it can be unreliable. When interpreting occasional pottery shards in a sediment body, some thought should be given as to their provenance and therefore their possible antiquity prior to incorporation into the sediment body. Abraded shards that have been transported downstream may well have been eroded from older sediments and can only ever yield a maximum possible date of the sediment. However, in floodplain and colluvial sequences, unabraded pottery is likely to have been incorporated into the sediment directly and is less likely to predate incorporation significantly. There also exists a serious sampling problem with pottery. One or two shards can easily give a false impression of antiquity, when a systematic search can provide a wide range of pottery ages (Brown 2009b). In this case, the youngest pottery can be used to provide a maximum age of deposition. If pottery cannot be dated stylistically, then it can be dated using thermoluminescence (see Chapter 6 for a description of this method). A pottery date derived in this way can then be compared with direct sediment dating using optically stimulated luminescence (Brown 1997).

Structures

Structures record a stable landsurface or, in the case of quays and docks, a relative sea level. House floors or foundations can provide a datum in an aggrading sedimentary sequence. Bridges are particularly useful since they record both ground level and the location of a channel, and also in some cases provide some idea of the size of the channel. This is exploited in one of the case studies described later in this chapter. In the Old World, Roman bridges are particularly common and can be used to estimate the post-Roman overbank deposition rate (Figure 3.1). One cautionary point is that some bridges, those with stone piers in or at the edge of the channel, do influence channel processes and so may lead to a biased view of river behaviour. Quays and docks can be related to a sea level, and when (as is commonly the case in the Mediterranean) they are now submerged or elevated, then relative sea level must have changed. A good example of this is the 600 m long jetty of the Roman port of Leptiminus in Tunisia, which now lies 0.6 m below sea level due to neotectonic activity (Brown *et al.* 2011). The inland preservation of old quays, docks and ports is some of the strongest evidence of high sedimentation rates in the Classical and post-Classical Mediterranean. Structures may also provide evidence of geological events with geomorphological implications. The most obvious is earthquake damage to ancient buildings. In some cases, the earthquake history can be linked to changes in drainage patterns (Jackson *et al.* 1996) and other geomorphic events such as landslides. Indeed, both neotectonic studies in the Old and New Worlds have frequently used archaeological evidence (Vita-Finzi 1988; Keller and Pinter 1996).

Many Old World rivers and increasingly rivers in the Americas and Australasia are so controlled by artificial banks, weirs and even roofs so that they have in effect become archaeological or historical structures themselves. Recent surveys along the Thames in London and along several French rivers have located thousands of structural remains such as wharfs, quays, revetments, weirs, bridges, culverts and even dwellings (Dumont *et al.* 2009; Thames Discovery Programme 2012) and this has practical consequences for river management both in terms of physical control or restoration and in terms of heritage conservation. The implication of structures, particularly weirs, for long-term system dynamics is only now being fully appreciated. Indeed, the classic view of channel form and floodplain morphology has been challenged by the proposition that for mid-Atlantic and western streams of the United States, form is largely a legacy of the impoundment of the valley floors

Table 3.1 Summary of archaeological tools in geomorphology.

Archaeological data	Geomorphic use	Advantages	Disadvantages
Pottery	Dating	Can be precise and the only dating evidence available	Can greatly overestimate the age or be unreliable
	Tracing	Can indicate potential sediment sources	Is more easily transported than the equivalent sediment size May have been reworked
Coins	Dating	Can be precise and the only dating evidence available	May have been reworked and/or in circulation for many years
Hearths	Dating	Can provide relatively pure carbon for ^{14}C dating Can provide information on the prevailing ecology	Must be in situ otherwise the date will be overestimated
	Alluviation rate	Provides a datum for calculation of the accumulation rate	Must be in situ otherwise the date will be overestimated
Bone	Dating	Extraction of collagen provides C for ^{14}C dating	May be transported and can be contaminated
Earthworks	Geomorphic stability	Can indicate age and conditions of a slope to limit estimates of landscape change	Slope evolution affected by the earthwork so location is not random
	Erosion rate estimation	Degradation or gullying of the feature can provide an estimation of erosion rate	The age and initial height must be known and the assumption made that the artificial slope is in some way representative of natural slopes
Middens	Shoreline location and sea-level estimation	Middens composed of discarded mollusca can indicate location of past shorelines of lakes or the sea and can be used to constrain height estimations	Must be accurately dated, which can be difficult May have been subjected to neotectonic effects
Land divisions: banks and walls	Erosion and colluviation rate estimation	Depths of sediment behind banks and walls can provide minimum estimate of within-field erosion and translocation	Wall or bank pre-dated accumulation of sediment Date of a wall or bank is often unknown
Stonework	Weathering rate estimation	Depth and character of stone weathering can be estimated in relation to areas of surface protection or metallic components	Cut stone weathering is related to natural weathering rates Age of stonework must be known Must assume no past re-cutting or cleaning of faces
Structures: buildings	Ground surface	Most buildings can be related to past surface level from steps doorways, etc. If buried, can indicate accumulation rate, or if elevated, an erosion rate	Date of building must be known and period that relates to ground surface. Some structures constructed below and others above ground level
Structures: bridges	Channel and ground level information	Indicates past channel location, can constrain estimates of channel width, depth and even discharge and can give contemporary floodplain height	Bridge may have altered local geomorphic rates Must assume there was no other channel on floodplain and bridge locations were representative of entire river reach
Structures: quays, wharves, jetties	Shoreline location and sea-level estimation	These structures can indicate past shorelines and provide evidence for past relative sea levels	Must be able to be dated Relationship to sea level (or tidal range) must be known or estimated Neotectonic effects must be recognized
Structures: wooden	Dating	May be able to provide a calendar date by ^{14}C or dendrochronological dating and this can allow the phasing of sedimentation with unparalleled temporal precision	Tree-ring dating requires suitable species (e.g. oak) and wood of sufficient diameter (15 cm minimum), wood can be reworked and re-used
Wood: tree rings	Palaeoclimatic reconstruction	Using the variation in ring width it is possible to reconstruct past growing season conditions particularly precipitation	Only possible for a few species (e.g. oak) and difficult to calibrate unless part of the chronology can be related to a documentary series (e.g. precipitation or streamflow record)
Structures: wells, cisterns, aqueducts, drains, etc.	Palaeohydrological estimation	Can provide indications of past groundwater levels and possibly surface water discharge	Many assumptions: both chronological and hydrological, for cisterns, aqueducts, drains, etc., it must be assumed that design Q was an accurate estimation of prevailing hydrological regime (see text)
Legacy sediment	Sedimentation rates; geomorphic impacts	Identifies stratigraphic location of pre-disturbance cultural evidence Integrates environmental impacts of cultural activities in watershed	Not preserved in all locations; evidence obscured by high water tables

Figure 3.1 The remains of a Roman bridge across the river Treia in Central Italy.

by water-powered mills (Walter and Merritts 2008). A similar anthropogenic trajectory can be seen in European rivers, although it took place over a much longer period and was driven initially by agriculturally induced sedimentation (Lewin 2010). The realization of this 'artefactualization' of lowland rivers has management implications in that sustainable restoration designed to maximize ecosystem services must be predicated on the awareness that such rivers are imprisoned in the banks of their history.

Palaeohydrological data from archaeology

Since many archaeological structures were originally designed to carry or store water under certain circumstances, their dimensions can be used to estimate some parameter of the past hydrological regime. Examples of such structures include dams, aqueducts, water supply structures (leets), wells, cisterns, connecting tunnels, channels for mining (ground sluicing) and drainage ditches. There are, however, significant methodological problems and convincing studies are rare. Assuming the structure can be accurately dated there remain several assumptions:

1 the function of the structure is in no doubt;
2 the design capacity is a reasonable indicator of prevailing hydrological conditions;
3 the full instantaneously functioning system is known (e.g. the number of functioning channels or pipes).

Dams have been used to estimate rainfall in arid regions associated with floodwater farming (Gilbertson 1986), and Gale and Hunt (1986) attempted to use floodwater farming structures in Libya to reconstruct water supply during the Roman period. They used the Darcy–Weisbach equation and an expression for roughness in turbulent flow in rough channels. Aqueducts, leets, and mining channels (for ground sluicing) can be used to calculate discharge and thereby estimate minimum precipitation. Drainage and artificial channels used in mineral processing should also reflect the prevailing hydrological conditions and water supply capacities of the system. Bradley (1990) attempted to use the dimensions and slope of tin streaming (alluvial mining) channels to estimate Late Medieval stream powers in southwest England and he used this estimate to support the observation that downstream floodplain and channel sediments were relatively enriched in cassiterite content of the >63 μm particle size fraction. Masonry drainage structures can also provide palaeohydrological data, an example being the hydraulic analysis by Ortloff and Crouch (1998) of a complex outlet structure in the Hellenistic city of Priene in modern Turkey, which provided evidence for a lower bound estimate of steady-state water supply to the city. Shallow wells have considerable potential to indicate past water-table height (or a minimum altitude), but a regionally based realization of this potential has yet to be attempted. One of the values of using archaeological data is that it provides paleoclimate information for regional water budgets, which is essential to understanding climate change.

Artefacts and fluvial processes

The characteristics of archaeological tracers and the deposits in which they occur may indicate important aspects of their source, mode of transport and age. Careful observations should be made to determine the condition of artefacts, whether they occur in primary positions of human deposition or in secondary deposits, and the geomorphic setting. For example, the amount of abrasion on individual artefacts may indicate distance from their source. Abrasion increases downstream, as has been shown with modern facsimiles of flint hand axes (Harding *et al.* 1987; Macklin 1995). Concentrations of tracer materials generally decrease with distance downstream due to dilution by barren sediment from local storage sites and from tributaries. The shape and density of artefacts may affect transport distance. For example, disk-shaped blades will travel further than more spheroidal axe and hammer heads, and bone will travel further than stone in a given flow environment. In addition to downstream travel distance, these factors may influence the lateral distance across floodplains that artefacts travel away from channels during large floods.

Mining sediment as tracers

The link between cultural activities and sedimentation is particularly well expressed by mining. Mining sediment not only provides evidence of fluvial processes, but also provides prime examples of fluvial responses to human alterations of the environment. All extractive mining and mineral processing produces some waste, which is either separated using rivers, is deliberately added to rivers, or eventually enters rivers via natural geomorphic processes. This line of geomorphological research can be traced back to the classic study by Gilbert (1917) of mining in the Sierra Nevada, which produced more than 10^9 m^3 of sediment (Gilbert 1917; James 1999). Several workers have distinguished between two types of sediment transport: *active transformation* where the fluvial system is transformed by the introduced waste (e.g. Gilbert's study) and *passive dispersal* where sediment markers are passed downstream mixed with the natural sediment without causing a substantial change in channel morphology (Lewin and Macklin 1987). Although useful, this distinction describes a continuum of possible conditions based on the degree of fluvial change and the potential to detect that change. The ability to distinguish mining sediment from other alluvia can be problematic. Mining is often accompanied by other land-use and river hydraulic changes, such as agricultural intensification, deforestation or construction of levees, so sediment loads may increase indirectly from other sources (Mossa and James 2013). Some of these questions are discussed in the case studies presented later.

Mining sediment is often studied because it forms distinctive stratigraphic units that can be recognized throughout a river course, dated, and related to specific cultures or activities. For example, the presence of slag material in floodplain deposits provides evidence of alluvium deposited after the 13th century as described in Case Study 3. Mining often amplifies background sediment loads by more than an order of magnitude, as was shown in a basin-wide analysis by Gilbert (1917) and demonstrated in a paired-watershed study of strip mining in Kentucky (Collier and Musser 1964; Meade *et al*. 1990). Several studies have documented severe alluvial sedimentation and channel morphological changes below mines in Great Britain (Lewin *et al.* 1977; Lewin and Macklin 1987) and North America (Gilbert 1917; Graf 1979; James 1989; Hilmes and Wohl 1995; Lecce and Pavlowsky 2001; Knox 2006).

Mining sediment is often rich in metals (Reece *et al.* 1978; Leenaers *et al.* 1988). The distinct signature of heavy metals associated with many mines often allows a local metal stratigraphy to be developed downstream of mines (Mossa and James 2013). For example, Knox (1987) was able to correlate floodplain strata with elevated concentrations of lead and zinc with periods of mining in southwest Wisconsin. Wolfenden and Lewin (1977) and Graf *et al.* (1991) developed similar chronologies for rivers in Wales and Arizona, respectively. Sediment sampling for evaluation of metals requires an understanding of fluvial transport processes and depositional environments. Heavy metals are often concentrated in the fine fraction of sediment due to sorting processes of the denser metalliferous particles, i.e. the principle of *hydrodynamic equivalency* (Rubey 1938). In mining sediment, however, the presence of multiple populations, including coarse metal particles, fine metal particles and coatings on or inclusions in particles of various sizes and densities, may complicate this relationship. The importance of particle coatings varies with the metals being sampled and ephemeral environmental factors such as pH, which encourage speciation into oxide, hydrous oxide and other phases. Most studies perform chemical analysis on a sand fraction isolated by sieving. Sampling and sieving should be performed with the minimum use of metal tools to avoid contamination. In a comparison of laboratory methods, Mantei *et al.* (1993) found that metal concentrations were homogeneous in the very fine sand grade, that splitting samples into quarters was not necessary and that crushing followed by sieving should not be done prior to chemical analysis.

Changes in metal concentrations below a source are often modelled as a simple downstream logarithmic decay function (Wertz 1949; Lewin *et al.* 1977; Wolfenden and Lewin 1978). Marcus (1987) showed that the downstream decay in copper was largely due to dilution by sediment from non-mining tributaries. Graf (1994) described the complexities involved in mapping downstream changes in plutonium and demonstrated a general decrease in concentrations downstream in tributary canyons to the Rio Grande. At the channel-reach scale, metal concentrations may vary greatly with geomorphic position. For example, Ladd *et al.* (1998) sampled 12 metals in seven morphological units of a cobble-bed stream in Montana. They found that concentrations varied between units; for example, eddy drop zones and attached bars had high concentrations whereas low- and high-gradient riffles and glides had low concentrations.

3.4 Legacy sediment

A growing need to recognize the impacts of past climate change and to separate these from changes caused by human activities has led to a need to understand potential links between sedimentation events and cultural activity. Thus, the traditional use of fluvial geomorphology as a tool in geoarchaeology is often reversed and archaeological evidence is used to interpret episodic sedimentation events. Where human activity was substantial, rarely can the effects of climate and anthropogenic change be completely separated on the basis of stratigraphic and sedimentological evidence alone. Knowledge of the location, timing and intensity of contemporary land use, along with an understanding of sediment processes, can, however, greatly constrain the possibilities and lead to a refined understanding of the interplay between climate and human activities. These concepts can be illustrated through the study of legacy sediment, that is, anthropogenic deposits generated by episodic erosion and sedimentation.

Evidence of environmental disturbance and fluvial adjustments

Episodic sedimentation generated by human land-use change, such as deforestation, ploughing for agriculture and mining, may be sufficiently severe to cause channel and floodplain aggradation that is preserved in the alluvial record. Aggradation is often followed by a period of recovery in response to relaxation of the causative factors (reforestation, cessation of mining, etc.) and channel incision that leaves sediment stored on floodplains. Together, the aggradation and incision have been referred to as an aggradation–degradation episode (ADE) (James and Lecce 2013) and the stored sediment that remains as legacy sediment. Although anthropogenic processes cannot be completely separated from climate-change processes, designation as legacy sediment implies an interpretation that anthropogenic processes played a substantial role in the sedimentation event. Hence some evidence of changes in land-use intensity should be demonstrated before defining deposits as legacy sediment. The presence of legacy sediment is an indicator of past cultural activity with a substantial environmental impact. It may be linked to land clearance or resource extraction and may signal substantial changes in population densities, land-use technology or socioeconomic changes such as the introduction of external markets for an export economy (Brierley *et al.* 2005). Recognition of legacy sediment is important to stream restoration and understanding fluvial processes. For example, much of classical fluvial theory from the mid-20th century was based on stream channels in the mid-Atlantic region of the United States; more recent work indicates that many of these streams were dominated by legacy sediment stored behind mill dams (Walter and Merritts 2008).

Legacy sediment – often referred to in different terms – is common in the Old World settings, where they may correspond to cultural events of great antiquity (Brown 1997; Lang *et al.* 2003; Hoffmann *et al.* 2008; Macklin and Lewin 2008) and in MesoAmerica (Beach *et al.* 2002). Legacy sediment in temperate North America and Australia is commonly associated with the relatively sudden introduction of agricultural, silvicultural and mining technologies from Western Europe and subsequent accelerated erosion and sedimentation (Knox 1972, 2006; Magilligan 1985; Lecce and Pavlowsky 2001).

The pristine New World myth

From the perspective of most fluvial geomorphologists working on alluvial stratigraphies in many areas of the New World, such as temperate North America and Australia, pre-Columbian land uses had limited impacts on sedimentation rates, often constrained to local deposits. This has led to a general concept in much of the fluvial geomorphology community in these areas that pre-Columbian cultures were not highly effective in generating sediment and that, in contrast, the arrival of European agriculture rapidly generated vast quantities of sediment. In contrast, however, many anthropologists and cultural geographers have challenged for decades the concept that pre-Columbian societies in the New World were environmentally benign and refer to this as the *Pristine Myth* (Denevan 1992, 2003; Butzer 1996; Redman 1999). Population densities in North America were much higher than previously estimated and substantial environmental changes were generated by these robust populations. These concepts, which represent an apparent contradiction between the common perception of many fluvial geomorphologists and many anthropologists, require attention. First, it is essential to differentiate between geomorphic and ecological impacts before addressing the question of environmental changes in the New World (James 2011, 2013). The impact of pre-Columbian societies on ecosystems was likely pervasive, but this is a separate issue from the relative geomorphic effectiveness

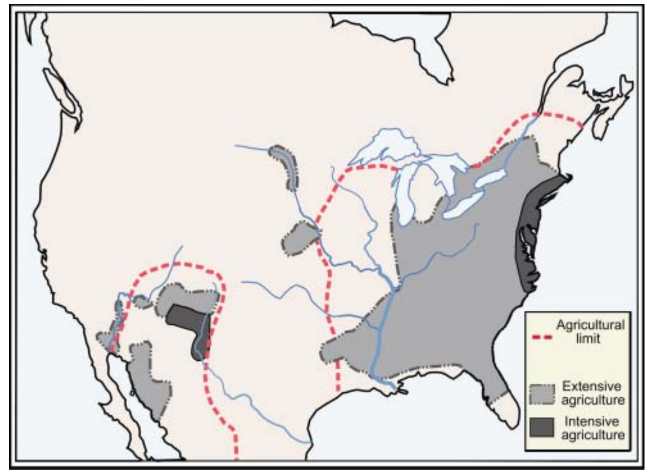


Figure 3.2 Pre-Columbian agriculture in North America. Adapted from James (2011), Denevan (1992; Extensive agriculture) and Butzer (1990; Intensive agriculture).

of pre-Columbian societies in the New World. Second, it would be folly to address this question without acknowledging geographic variations in cultural activity and landscape sensitivity. Maps of pre-Columbian agricultural intensity presented by Butzer (1990) and Denevan (1992) and adapted for North America by James (2011) show strongly non-uniform spatial patterns of intensive agriculture indicating that pre-Columbian sedimentation is highly unlikely in many areas (Figure 3.2). Some studies have documented local pre-Columbian alluvial deposits, but many more such studies are needed to constrain the spatial and temporal distribution of these deposits. As a counterpoint to this debate, Butzer (1996) has argued that the assumption of pervasive post-Columbian legacy sediment is ill-founded. He notes that European settlement, especially by Spaniards in MesoAmerica, was often accompanied by a strong land-use ethic and relatively little sedimentation over the last four centuries. In general, legacy sediment can be an extremely useful tool for recognizing cultural horizons in alluvial stratigraphies, computing sedimentation rates and identifying links between cultural activity and environmental impacts. Simple scenarios of land-use change and aggradation–degradation episodes should not be assumed, however, and a serious effort should be made to understand pre-existing conditions and the nature of perturbations leading to sedimentation. More work is needed on analysing the sediment underlying legacy sediment contacts.

3.5 Using archaeological data: case studies

In this part of the chapter, a series of case studies are used to illustrate how archaeological data can be used to answer geomorphological questions. The examples are taken from different climatic regions and cover different time-scales.

Case study 1. Fluvial reconstruction from bridge structures on the River Trent, UK

Archaeological finds can provide both the opportunity and raison d'être for the reconstruction of past geomorphic and hydraulic conditions. The exploitation of aggregate from large areas of the Middle Trent floodplain in central England has allowed the excavation and recording of hundreds of archaeological finds, including human and animal skeletal remains, log-boats, fish weirs, anchor weights, revetments, several bridges and a mill. Together with 'natural' finds such as tree-trunks, organic palaeochannel sediments and flood debris, this has allowed a geomorphological reconstruction of the Holocene evolution of the Middle Trent floodplain based on both an archaeological and radiocarbon chronology. From Hemington and surrounding investigations, a partial fluvial history is postulated in Table 3.2.

The Middle Trent has been characterized by channel change throughout the Holocene. The absence of a high slope (Hemington–Sawley average only $0.0006\,\mathrm{m\,m^{-1}}$) is most likely due to a rapid downstream increase in discharge from the four major tributaries that enter the main channel in under 40 km, the flood characteristics of two of these tributaries and an abundant supply of unconsolidated or cemented sandy gravels provided by the low and wide Devensian terraces. The unusual width of the Late Devensian (OIS 3–2) gravel terraces here is due to proximity to the Devensian ice margin, which was less than 30 km upstream. The early to mid-Holocene data are largely derived from palaeochannels whilst the late Holocene period is known in most detail due to the occurrence of buried bridges (Figure 3.3), a mill and weir and abundant other evidence of channel change. Several geomorphologists have attempted to use archaeological structures to quantify geomorphological parameters. This is based on the rationalist assumption that structures such as bridges, weirs, etc., were built to contain a certain flow, and functioned by containing a run of flows. In some cases, destruction of the structure by a flood can also be used to estimate the magnitude of the event. Geomorphological studies of three Medieval bridges buried under gravels in the floodplain of the Middle Trent have employed several of these techniques, including simple slope–area calculations of discharge from channel dimensions, HEC-II flow modelling and palaeohydraulic calculations based upon transported and non-transported clasts (Brown 2009c, 2011b). The sedimentology, the archaeology and the pattern of palaeochannel fragments suggest that this reach of the Middle Trent was highly unstable during the Holocene and especially the last 1000 years. The sedimentology suggests relatively shallow, unstable channels eroding and depositing sand and gravel. The predominant sedimentological features, horizontal and low-angle bedding with shallow channels, suggests a locally

Table 3.2 A chronology of the channel change Hemington–Sawley reach of the Middle Trent derived largely from geoarchaeological studies.

Period	Channel type	Sites	Notes
Windermere Interstadial	Meandering	*Hemington*, basal channel peat (Brown 2008)	Down-cutting into terraces and bedrock
Loch Lomond Readvance	Braided	*Hemington* basal gravels (Brown 2008), *Church Wilne* (Coope and Jones 1977; Jones *et al*. 1977), *Attenborough* (BGS, A.G. Brown unpublished)	Deposition of basal 'Devensian' gravels and intense frost action creating polygons
Mesolithic	Low sinuosity, possibly multiple-channel (anastomosing)	*Shardlow*-stocking palaeochannels (Challis 1992; Knight and Howard 1994), *Repton* (Greenwood and Large 1992), *A6 Derby By-pass* (A.G. Brown unpublished), *Attenborough* (BGS, A.G. Brown unpublished)	Some avulsion leaving linear palaeochannels, which are often over 1 km from the present channel
Neolithic	Multiple channel-braided, low sinuosity	*Hemington* (Clay and Salisbury 1990), *Colwick* (Salisbury *et al.* 1984), *Langford* and *Besthorpe* (Knight and Howard 1994)	Fishweirs and black oaks in small, shallow channels
Bronze Age	Meandering?	*Colwick* (Salisbury *et al*. 1984), *Collingham* (M. T. Greenwood, personal communication)	Little evidence except at Colwick and downstream
Iron Age and Roman	Meandering, sinuous, point-bar sediments	*Holme Pierrepoint* (Cummins and Rundell 1969)	Palaeochannel associated with settlement at Sawley and evidence of settlement on the terraces, excavated site RB site at Breaston (Todd 1973)
6th–9th centuries AD	Meandering, highly sinuous	*Hemington* (Brown *et al.* 2010)	Large palaeochannel dated by radiocarbon and palaeomagnetics
11th–13th centuries AD	Braided, unstable	*Hemington*, *Colwick* (Salisbury *et al.* 1984), *Sawley* palaeochannel	Channels associated with the bridges
17th–19th centuries AD	Anastomosing to single channel, moderate to low sinuosity	*Hemington*	Avulsion sometime between 15th and 17th centuries from the Old Trent to the modern Trent
19th–21st centuries AD	Meandering, stabilized	Map and documentary evidence	Embanked, partially regulated and engineered, construction of the Trent and Mersey canal, Sawley cut and Beeston canal

Figure 3.3 Photograph of last (13th century) Medieval bridge after excavation at Hemington showing a bridge pier and baffle at the upstream end. Source: Richard Buckley, Director, University of Leicester Archaeological Services.

braided river, which is in agreement with the low sinuosity of the channels during the early Medieval period. However, the preservation of old palaeochannels and archaeological features (such as the mill and bridges) and the avulsion of the channel sometime between the 15th and 16th/17th centuries suggest that the river underwent a braided and anastomosing phase before returning to a single-channel meandering form. The typical form of both braided and anastomosing reaches has been used in a generalized geomorphological model of the reach from the 8th to 19th centuries (Figure 3.4).

It is impossible to accommodate both the archaeology and the palaeochannels unless the reach has at least two (preferably three) functioning channels and the evidence suggests that there was a migration of channels eastwards, leaving a Prehistoric meander core, but that a functioning westerly channel remained (even if small) until an avulsion sometime between the 15th and 16th/17th centuries led to the abandonment of the easterly channel (Old Trent) and conversion of the westerly channel into the only permanent channel in the reach. Associated with this change in channel numbers is a drop in main channel sinuosity, as would be expected during a period of braiding and high bedload movement through the reach. This model, therefore, suggests that this reach of the Trent went from being a meandering single-channel river in the early Medieval period (6th–9th centuries) to a braided river in the 10th–11th centuries back to a single-thread meandering river by the end of the 17th century, probably passing through an early wandering-gravel bed phase and a later transitional anastomosing phase. This is a classic example of medium- to long-term metamorphosis of a river channel and floodplain. The processes responsible for this change are large floods, particularly those generated in the Pennine uplands, and an increase in the transport of bedload into and through the reach. The trigger for this remains unclear but may have been floods, probably rain on snow, that occurred during the 11th–13th centuries, a period that has been labelled the 'Crusader cold period' and is part of the Late Medieval Climatic Deterioration. The cycle of channel change is clearly related to abundant bedload supply and high sediment transport rates and can be viewed as channel adjustment to a pulse of sediment which was, through channel metamorphosis, deposited into floodplain storage. The nature of the reach (shallow channels) was taken advantage of for the construction of bridges, the builders presumably being unaware of the transitory nature of the channel conditions or constrained by the geography of the route. The climate changes of the Late Medieval period and early modern period are now considered probably to have been the most dramatic in the Holocene (Rumsby and Macklin 1996) and the Middle Trent is particularly sensitive to changes in hydrometeorology. This is not to say that there were no human impacts on these events, as deforested uplands are far more likely to produce large rain on snow events due to the increased depth of the snowpack that can accumulate over grass as opposed to tree cover. Likewise, there is little doubt that runoff generation times have been decreased, and therefore peak flows increased, by drainage and land-use change (Higgs 1987).

Bridges provide the most obvious evidence of channel change in the case of either bridges over palaeochannels or old bridges over modern channels. There is, of course, a conceptual problem with channel evidence from bridges, because they may not be randomly located along channels, most replaced fords, and in many cases were clearly located where the channel and floodplain were constricted and a terrace or high bank could be used in construction. This will also depend upon the state of bridge technology, early bridges with restricted single spans were restricted to narrow or divided channels and later bridges requiring solid foundations on terraces or bedrock. However, the geographical location of a bridge depends upon population patterns, with routes linking towns or villages by the shortest or most practical route. Thus bridge *site* and geomorphic history are fundamentally geomorphologically controlled (or unavoidable) whereas bridge *location* is generally dictated by routes linking centres of population, at least in lowlands. It has

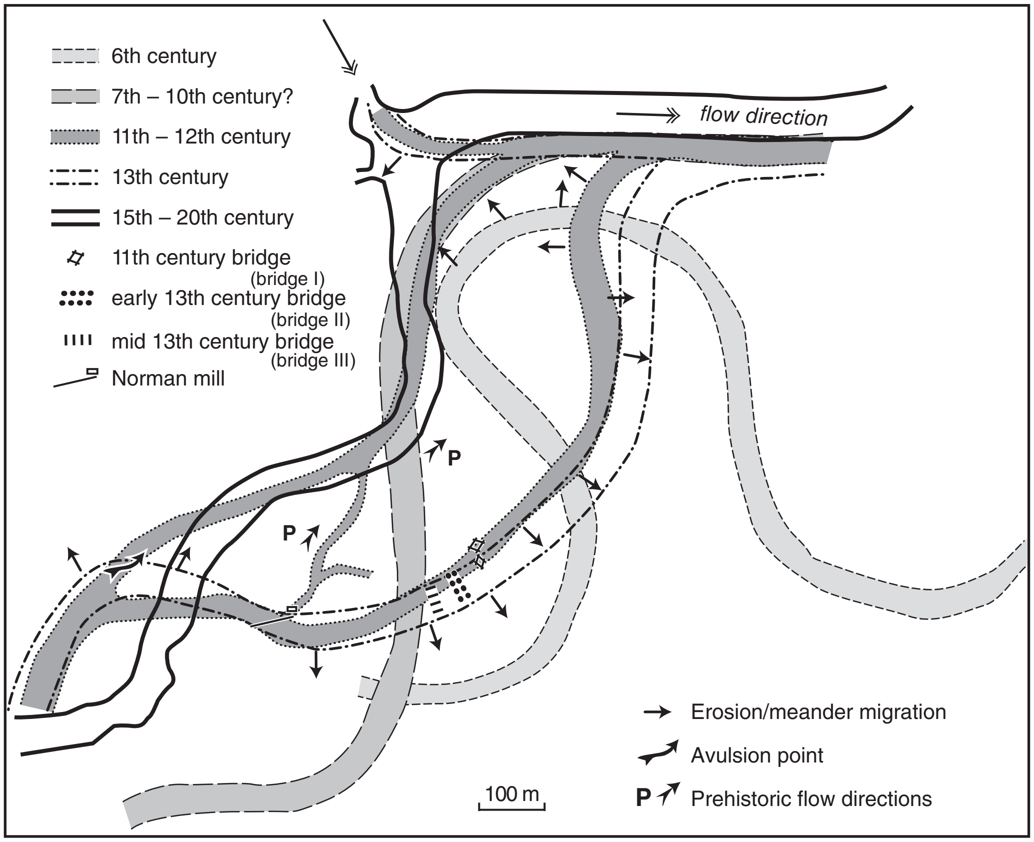


Figure 3.4 A model of channel change in the River Trent in the Hemington reach over the last 1000 years. The location and dating of each channel position are derived from archaeological data including the Medieval bridges, fishweirs, a mill and anchor stones and also sedimentological data. From Brown *et al*. (2013).

also been argued that bridges prevent channel change; although this is certainly true in the case of lateral migration, where it depends upon continued capital investment in the structure, it is not true where avulsion is a major cause of channel change.

Case study 2. Slags, bedload and hydraulic sorting in Belgium

Bedload progression has been evaluated in rivers using slags coming from old ironworks settled in the south Ardennes valleys at the early 17th century (Sluse and Petit 1998). During these periods, the slags were disposed of into the rivers. They are still being transported even if the factories have been closed for a considerable period of time. The slags are easily recognizable thanks to their visual characteristics. Their average density is 2.1. The slags have been sampled in 19 riffles situated along the River Rulles, in its tributaries where ironworks have been installed, and downstream, in the River Semois into which the River Rulles flows. Figure 3.5(a) shows the trend of the trend of D_{90}, that is, the size corresponding to of the tenth percentile on the frequency distribution (10% are coarser), along the River Rulles course, using a cumulative distance from the most downstream of the iron factories (explaining the decrease in size in sites 1–3). The slags brought down by tributaries explain the increase in the slag size in the Rulles (examples: site 4 and sites 7–10). Slags have also been found in the River Semois (site 15) but none 4 km downstream of this last site.

A relationship is drawn between the slag size and the distance from the ironworks where these slags have been discarded into the river (Figure 3.5). This curve shows a rough decrease in particle size, which decreases from 80 to 20–30 mm in diameter in less than 5 km; subsequently the slag size decreases only slowly. The slag refining in the first few kilometres downstream of the ironworks does not result from modifications in hydraulic characteristics of the river or a diminution of its competence. Indeed, the unit stream powers remain identical along its course. This slag size reduction does not result from abrasion, from granular disintegration or from gelifraction effects (Sluse and Petit 1998). It results from a hydraulic sorting occurring in the first few kilometres downstream of the input sites.

The slag size, which, after 5 km, remains almost constant regardless of the distance, represents the actual competence of the river (the particle size transported along substantial distances and evacuated out of the catchment). The particle size (12 mm maximum with regard to equivalent diameters using a density of 2.65) is relatively small, but is justified by the low values of unit stream power (25–30 W m^{-2} at the bankfull discharge). Higher competence causes the hydraulic sorting, but this is exerted only locally and during intense events.

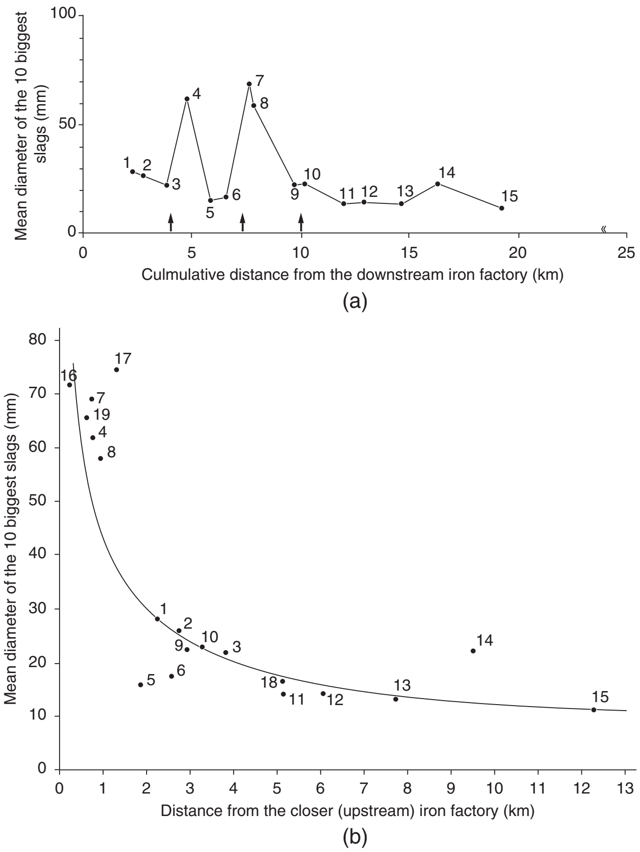


Figure 3.5 (a) Trend of the mean diameter of the 10 biggest slags measured by the *b*-axis, along the River Rulles and the River Semois, using a cumulative distance from the most downstream iron foundry located on the Rulles. The arrows on the *x*-axis indicate the junctions of the tributaries where ironworks were located. The star symbol on the *x*-axis indicate the upstream limit of slag deposition in the Semois. (b) The diameter of the 10 biggest slags measured by their *b*-axes in relation to distance from the closest iron foundry. Source: Sluse and Petit, 1998. Reproduced with permission of Geographie Physiques et Quaternaire.

Several slags (10–14 mm in diameter or 9–12 mm using equivalent diameters) have been found 12.5 km downstream of the closer iron factory, which produces a bedload wave progression of 3.3 km per century (Figure 3.6a). The most upstream site in the River Semois where no slag has been found shows that the bedload wave progression is less than 17 km since the middle of the 17th century (<3.9 km per century). Such progression is low in comparison with other studies (between 10 and 20 km per century), although most of those were of mountain rivers with strong energy (Tricart and Vogt 1967; Salvador 1991).

Case study 3. Artefactual evidence of floodplain deposition and erosion in Belgium

The rate of floodplain formation has been estimated in Ardennes rivers using stratigraphic markers identified by Henrottay (1973). These consist of scoria (smaller than 105 μm) produced by the medieval metal industry set up in Ardennes valleys from the mid-13th century. The debris from these factories

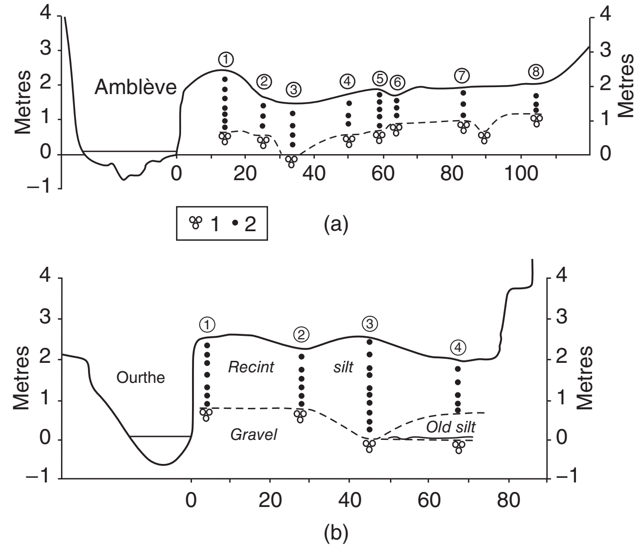


Figure 3.6 Transverse profiles of (a) the River Amblève and (b) the River Ourthe with the presence of microscopic scoria in the floodplain fill, the depths of which provides an estimation of the sedimentation rates since the 13th century. Key: (1) gravel and (2) silt with scoria. Circled numbers 1–8: cores.

was dumped into the rivers so that the presence of microscopic scoria in alluvial deposits affirms that the floodplain was built after the 13th century. As shown by Figure 3.6(a), concerning the River Amblève, the whole floodplain contains microscopic scoria deposited after the 13th century. The thickness of recent flood silt generally exceeds 1 m and frequently reaches 2 m, which gives a rate of accumulation of 28 cm per century. Henrottay prospected different rivers of the Ourthe basin and the River Meuse downstream of Liège (Table 3.3). The rate of sedimentation generally exceeds 20 cm per century. Everywhere the thickness of silt deposited since the 13th century is greater than the layer of old silt deposited prior to scoria deposition. Human

Table 3.3 Sedimentation and erosion rates determined using microscopic scoria deposited in floodplain sediments.

River	Catchment area (km^2)	Date of ironworks	Sedimentation rate (cm per century)	Lateral erosion rate (m per century)
North Ardenne (from Henrottay 1973)				
Amblève	1044	1250	23.5	14.6
Ourthe	1597	1250	28–33	6.3
Ourthe	2691	1250	28	—
Somme	38	1400	8–18	3.9
Meuse	802	1250	21	42
South Ardenne (from Sluse 1996)				
Rulles	96	1540	14.4	5.5
Rulles	134	1540	9.1(6)	4.4
Mellier	63	1620	19.6(5)	5.4
Rulles	220	1540	24.9(5)	18.0
Semois	378	1540	19.8(5)	33.0

activities (deforestation and expansion of the area under tillage) have probably played a dominant role in the silt accumulations in the valleys. The same technique was used in the south of the Ardenne by Sluse (1996). The rates of sedimentation are slightly less than in the north of the Ardenne (Table 3.3). Two reasons explain this difference. Deforestation in the south Ardenne catchments is now less important and the present land use of these watersheds is dominated by forests and pastures, so that soil erosion is less than in the north part of the Ardenne. Furthermore, the loess deposits are less thick in the south Ardenne and there is therefore less material to erode.

The microscories allow the evaluation of the importance of lateral erosion of these rivers (Henrottay 1973). As shown in Figure 3.6(a), the silt contains microscopic scoria and was therefore deposited after the middle of the 13th century, along all the width of the floodplain. Silt without scoria (before the 13th century) has been eroded, which implies that from that time the river has swept away, at least once, all of its floodplain across a width of 100 m. This gives valuable indications of lateral erosion rates. In this case, it achieved an average rate close to 15 m per century. In contrast to the River Amblève, the Ourthe has not systematically swept the totality of its floodplain since one can find old silt on which rests the recent silt (Figure 3.6b). This nevertheless documents lateral erosion of at least 45 m. The rates of lateral erosion are similar in south Ardenne rivers (Table 3.3). Using this method, it is clear that the lateral erosion can be underestimated because the river may have passed the zone where the old silt was eroded several times and this may explain low lateral erosion values. However, the rates agree with measurements taken from old maps (Petit 1995).

Case study 4. Metal mining and fluvial response: in the Old and New Worlds

Tin mining, which produces large amounts of sediment, has a long history, since it is one of the constituents of bronze and has been mined in Europe since the beginning of the so-called Bronze Age (third millennium BC in Great Britain). Both archaeologists and geomorphologists have a shared interest in the period before written records – the archaeologist in using sediments to search for pre-Medieval tin mining and geomorphologists in both dating alluvial deposits and understanding river behaviour on the 10^3 years time-scale. A geochemical survey of rivers draining Dartmoor, southwest England, was undertaken in order to address both of these questions (Thorndycraft *et al.* 1999, 2004). In this case, archaeological evidence of pre-Medieval tin mining is unlikely owing to the almost complete reworking of any earlier deposits by late- and post-Medieval tin mining and streaming. Floodplain sedimentary successions, which had not themselves been mined but are downstream of known areas of tin streaming, were found to retain a geochemical record of the mining activities because the early tin streaming released large quantities of mine-waste tailings. Radiocarbon dating of these sequences has shown an excellent match with the documentary record, confirming a first phase of streaming commencing in the 12–13th centuries, reaching a maximum in the 16th century and a later phase in the 19th and early 20th centuries (Figure 3.7). A combination of XRF on particle size fractionated sediment and SEM/EDS studies of density separated samples allowed the geochemical characterization of, and distinction between, streaming waste and naturally tin-enhanced sediments. In the Avon, Teign and Erme valleys in southwest England there is considerable overbank sediment aggradation coupled with the tin enhancement, and this was probably associated with changes in channel pattern and morphology.

Another example of the use of archaeological/historical data in fluvial geomorphology is the study by James (1989) of hydraulic gold mining sediments in the Bear River, California. In the lower Bear Basin, subsurface coring indicated that about 106 million m^3 of mining sediment remained stored 100 years after the cessation of gold mining. This estimate was more than double previous estimates and indicated that over 90% of the lower basin deposits remain in storage. Both topographic and historical evidence was used to illustrate the continued reworking of mining sediment as relatively frequent flows are competent to move channel-bed material derived from mining sediment. As sediment loads are still greater than pre-mining values, in contrast to Gilbert's (1917) symmetrical wave model of geomorphic response based on a rapid return of channel-bed elevations to pre-mining values. This suggests that the empirical foundation of the symmetrical wave model is biased and that a distinction should be made between bed waves based on bed elevations and sediment waves representing sediment flux (James 2006, 2010). Channel incision and hence sustained erosion and deposition in the Bear River have been promoted as prolonged reworking of stored sediment that is governed by several factors in addition to decreased sediment supply, probably a function of catchment and valley topography and geomorphic conditions.

Another clear example of active transformation and the persistence of anthropogenic sediment, in this case associated with tin mining, is the work by Knighton (1989, 1991) on the Ringarooma basin in Tasmania. Mining in the basin lasted for over 100 years from 1875 to 1982, during which time 40 million m^3 of sediment was added to the river. The result was channel metamorphosis with bed aggradation, an increase in width where the channel was not confined and the development of a multiple channel pattern. Only now is degradation in the upstream reaches returning the river to something like its pre-mining condition. Similar results have come from studies of the fluvial response to lead mining in upland Britain and in particular the combined effects of increased sediment supply and climate change in the form of perturbations in the flood frequency/magnitude (Macklin *et al.* 1992; Hudson-Edwards *et al*. 1999).

Both the archaeological and historical studies of mining and other legacy sediments in the Old and New Worlds lead to two geomorphic conclusions. First, the Gilbert symmetrical

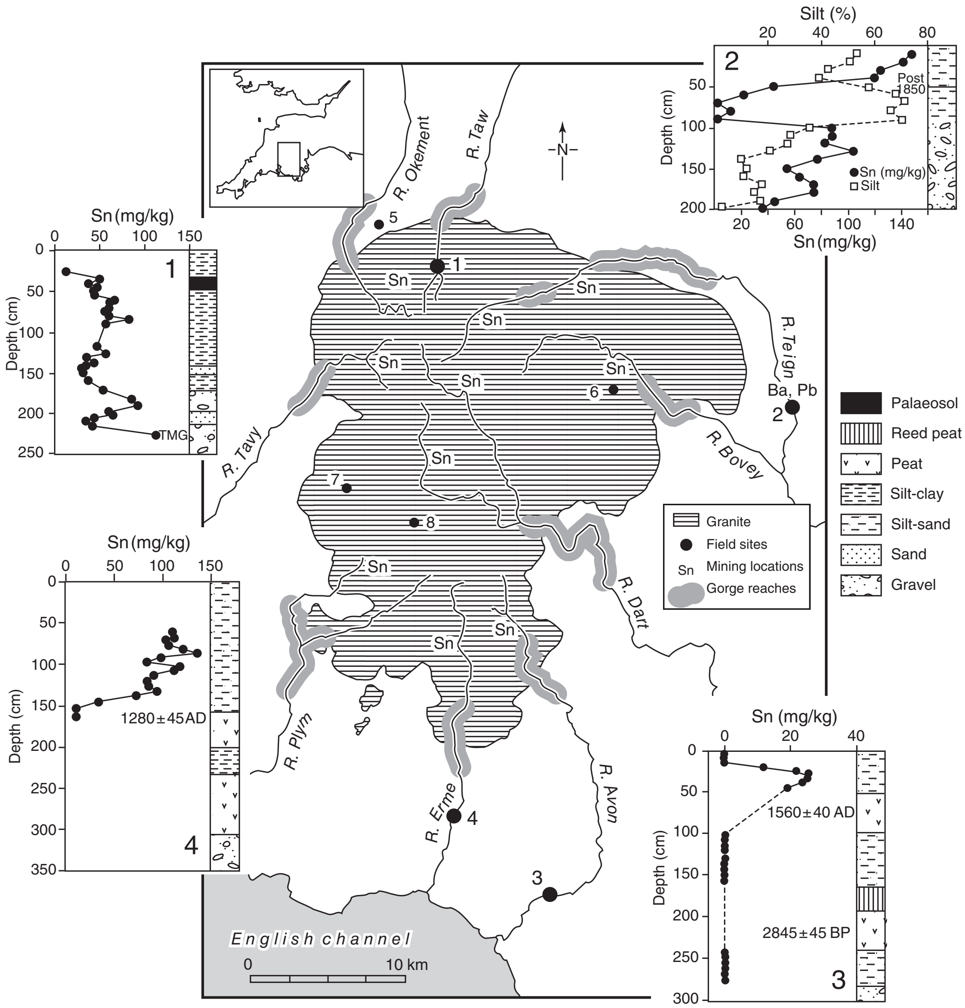


Figure 3.7 The distribution of Medieval tin mining sites (tin streaming sites) and geochemical profiles of alluvial sections in the floodplains of the rivers draining Dartmoor in southwest England.

wave model and river response is itself a function of basin conditions including basin topography, channel pattern and long-term geomorphic trends, such as neotectonically induced incision/aggradation, in addition to post mining flood history. Second, in basins where storage goes overbank, the residence time of episodically derived anthropogenic sediments can be on the order of geological time, being on millennial rather than decennial or centennial time-scales. This finding has important implications for the release of stored contaminants from floodplains in response to changing forcing conditions such as global warming.

3.6 Conclusions

Archaeology can provide far more valuable information than just dating. Indeed, to some extent, dating has now become the prerogative of the geomorphologists with artefact typological chronologies being re-evaluated as a result of the development of sediment-based dating techniques (luminescence and cosmogenic radionuclides). Archaeology can provide rapid evidence of land surfaces, sediment reworking and palaeoenvironmental conditions. It can also, under favourable conditions, set parameters that can be used in the modelling of past processes. hence archaeological data – including artefacts and the methods developed for their study – have led to the development of a set of tools that can be used by geomorphologists to study past fluvial processes and hydrological change. Conversely, fluvial geomorphology provides a series of tools that, if properly understood and applied, can be used to study both the timing and environmental context of cultural impact on the landscape (Howard and Macklin 1999; Brown 2008). The explanation for such strong linkages between archaeological and geomorphic methods arises from interactions between human societies and fluvial landforms, that is, river channels and floodplains. These interactions include anthropogenic alterations of fluvial processes and magnitude–frequency relationships in addition to the incorporation of human relics in alluvium. This interaction is exemplified by the artefactualization of rivers and the geomorphological implications that result.

In Europe, Asia and Africa, substantial anthropogenic environmental disruptions began in the Middle Holocene and the clear cultural record allows the application of these tools over a relatively long period. In the Americas and Australia, the early cultural record can be more subtle and extensive agriculture and deforestation often came much later, leaving an abrupt boundary late in the stratigraphic record. There are advantages to both situations. In the Old World, we can learn about the effects of long-term and multiple intermittent anthropogenic perturbations, and in the New World, we can study the effects of the sudden introduction of environmental exploitation (e.g. the geomorphic response to mining). Both of these lessons are essential to an understanding of the future potential for human impacts on the environment and global environmental changes. It is unfortunate that one of the driving forces of increasing links between archaeologists and geomorphologists has been the relentless drift of funding towards applied and short time-scale studies in geomorphology. Although such process-oriented studies are important, they cannot replace the need for an empirically based understanding of Earth-surface processes over millennial time-scales.

References

Antevs, E.V. 1935. Age of the Clovis Lake beds. *Proceedings of the Academy of Natural Sciences of Philadelphia* 87: 304–312.

Antevs, E.V. 1955. Geologic–climatic dating in the West. *American Antiquity* 20: 317–335.

Baker, V.R., Kochel, R.C. and Polach, H.A. 1985. Radiocarbon dating of flood events, Katherine Gorge, Northern Territory, Australia. *Geology* 13: 344–347.

Beach, T., Luzzadder-Beach, S., Dunning, N., Hageman, J. and Lohse, J. 2002. Upland agriculture in the Maya Lowlands: ancient Maya soil conservation in north western Belize. *The Geographical Review* 92: 372–397.

Bradley, S.B. 1990. Characteristics of tin-streaming channels on Dartmoor, UK. *Geoarchaeology* 5: 29–41.

Brierley, G.J., Brooks, A.P., Fryirs, K.A. and Taylor, M.A. 2005. Did humid–temperate rivers in the Old and New Worlds respond differently to clearance of riparian vegetation and removal of woody debris? *Progress in Physical Geography* 29: 27–49.

Brown, A.G. 1997. *Alluvial Geoarchaeology: Floodplain Archaeology and Environmental Change*, Cambridge: Cambridge University Press.

Brown, A.G. 2000. Geomorphological techniques in Mediterranean landscape archaeology. In: Barker, G. and Mattingly, D., eds., *The Archaeology of the Mediterranean Landscape 2: Environmental Reconstruction in Mediterranean Landscape Archaeology*, Oxford: Oxbow Books, pp. 45–54.

Brown, A.G. 2008. Geoarchaeology, the four dimensional (4D) fluvial matrix and climatic causality. *Geomorphology* 101: 278–297.

Brown, A.G. 2009a. *Aggregate-related Archaeology: Past, Present and Future*. Kings Lynn: Heritage Marketing and Publications/Oxford: Oxbow Books, 220pp.

Brown, A.G. 2009b. Colluvial and alluvial response to land use and climate change in Midland England: an integrated geoarchaeological approach. *Geomorphology* 108: 92–106.

Brown, A.G. 2009c. The geomorphology and environment of the Hemington reach. In: Ripper, S. and Cooper, L.P., eds., *The Hemington Bridges. The Excavation of Three Medieval Bridges at Hemington Quarry Near Castle Donington, Leicestershire, Leicestershire Archaeology Monograph 16*, Leicester: University of Leicester Archaeological Services, pp. 96–113.

Brown, A.G. 2011a. Dating of surfaces and sediments. In: Gregory, K.J. and Goudie, A.S., eds., *SAGE Handbook of Geomorphology*, London: SAGE PPublications, pp. 192–209.

Brown, A.G. 2011b. Palaeohydrology, climate change and bridges. In Prell, M., ed., *Archäologie der Brücken: Archaeology of Bridges*, Regensburg: Verlag Friedrich Pustet, pp. 214–220.

Brown, A.G., Cooper, L., Salisbury, C.R. and Smith, D.N. 2001. Late Holocene channel changes of the Middle Trent: channel response to a thousand year flood record. *Geomorphology* 39: 69–82.

Brown, A.G., Ellis, C. and Roseff, R. 2010. Holocene sulphur-rich palaeochannel sediments: diagenetic conditions, magnetic properties and archaeological implications.*Journal of Archaeological Science* 37: 21–29.

Brown, A.G., Sherriff, B. and Butler, S. 2011. Environment. In: Stone, D.L., Mattingley, D.J. and Ben Lazreg, N., eds., *Leptiminus (Lamta). Report No. 3. The Field Survey*, Portsmouth, Rhode Island: Journal of Roman Archaeology Supplementary Series No. 87, pp. 29–48.

Brown, A.G., Carey, C.J., Howard, A., Toms, P.S. and Challis, K. 2013. A geological model of a floodplain confluence Central England, UK using LiDAR, geophysics (GPR & ER) and multi-method dating. *Earth Surface Processes and Landforms* 38:237–249.

Brown, A.G., Tooth, S., Bullard, J.E., Thomas, D.S.G., Chiverrell, R.C., Plater, A.J., Murton, J., Thorndycraft, V.R., Tarolli, P., Rose, J., Wainright, J., Downs, P. 2016. The Geomorphology of The Anthropocene: Emergence, Status and Implications. *Earth Surface Processes and Landforms*.

Butzer, K.W. 1990. The Indian legacy in the American landscape. In: Conzen, M.P., ed., *The Making of the American Landscape*, Boston, MA: Unwin Hyman, pp. 27–50.

Butzer, K.W. 1996. Ecology in the long view: settlement histories, agrosystemic strategies, and ecological performance. *Journal of Field Archaeology* 23(2): 141–150.

Challis, K. 1992. *Archaeological Investigations of a Proposed Stocking Area for Aggregate at Shardlow, Derbyshire*, Nottingham: Trent and Peak Archaeological Trust.

Clay, P. and Salisbury, C.R. 1990. A Norman mill dam and other sites at Hemington fields, Castle Donington, Leicestershire. *The Archaeological Journal* 147: 276–307.

Coles, B. and Coles, J. 1989. *People of the Wetlands: Bogs, Bodies and Lake-Dwellers*, London: Thames and Hudson.

Collier, C.R. and Musser, J.J. 1964. Sedimentation. In: *Influences of Strip Mining on the Hydrologic Environment of Parts of Beaver Creek Basin, Kentucky, 1955–59, US Geological Survey Professional Paper 427-B*, Washington, DC: US government Printing Office, pp. 48–64.

Coope, G.R. and Jones, P.J. 1977. Church Wilne. In: *The English Midlands*. International Quaternary Union Guidebook for Excursion A2, p. 25.

Cummins, W.A. and Rundle, A.J. 1969. The geological environment of the dug-out canoes from Holme Pierrepoint, Nottinghamshire. *Mercian Geologist* 3: 177–188.

Denevan, W.M. 1992. The pristine myth: the landscape of the Americas in 1492. *Annals of the Association of American Geographers* 82(3): 369–385.

Denevan, W.M. 2003. *Cultivated Landscapes of Native Amazonia and the Andes*. New York: Oxford University Press, 396pp.

Dumont, A., Moyat, P. and Verger, S. 2009. Dossiers d'archaeologie fluviale en Europe. *Sommaire* 331: 2–88.

Ellis, C.E. and Brown, A.G. 1999. Alluvial micro-fabrics, anisotropy of magnetic susceptibility and overbank processes. In: Brown, A.G. and Quine, T.A., eds., *Fluvial Processes and Environmental Change*, Chichester: John Wiley & Sons, pp. 181–206.

Gale, S.J. and Hunt, C.O. 1986. The hydrological characteristics of a floodwater farming system. *Applied Geography* 6: 33–42.

Gilbert, G.K. 1917. *Hydraulic-mining Debris in the Sierra Nevada, US Geological Survey Professional Paper 105*, Washington, DC: US Government Printing Office.

Gilbertson, D.D. 1986. Runoff (floodwater) farming and the rural water supply in arid lands. *Applied Geography* 6: 5–12.

Graf, W.L. 1979. Mining and channel response. *Annals of the Association of Americam Geographers* 69(2): 262–275.

Graf, W.L. 1994. *Plutonium and the Rio Grande: Environmental Change and Contamination in the Nuclear Age*. Oxford: Oxford University Press.

Graf, W.L., Clark, S.L., Kammerer, M.T., Lehman, T., Randall, K. and Schröeder, R. 1991. Geomorphology of heavy metals in the sediments of Queen Creek, Arizona, U.S.A. *Catena* 18: 567–582.

Greenwood, M.T. and Large, A.R.G. 1992. Insect sub-fossils from the Trent floodplain near Repton. *Journal of the Entomological Society* 107: 12–17.

Harding, P., Gibbard, P.L., Lewin, J., Macklin, M.G. and Moss, E.H. 1987. The transport and abrasion of flint handaxes in a gravel-bed river. In: Sieveking, G. de G. and Newcomer, M.H., eds., *The Human Uses of Flint and Chert*, Cambridge: Cambridge University Press, pp. 115–126.

Henrottay, J. 1973. La sédimentation de quelques rivières belges au cours des sept derniers siècles. *Bulletin de la Société Géographique de Liège* 9: 101–115.

Herz, N. and Garrison, E.G. 1998. *Geological Methods for Archaeology*, Oxford: Oxford University Press.

Higgs, G. 1987. Environmental change and hydrological response: flooding in the Upper Severn catchment. In: Gregory, K.J., Lewin, J. and Thornes, J.B., eds., *Palaeohydrology in Practice*, Chichester: John Wiley & Sons, pp. 131–159.

Hilmes, M.M. and Wohl, E.E. 1995. Changes in channel morphology associated with placer mining. *Physical Geography* 16: 223–242.

Hoffmann, T., Lang, A. and Dikau, R. 2008. Holocene river activity: analysing ^{14}C-dated fluvial and colluvial sediments from Germany. *Quaternary Science Reviews* 27: 2031–2040.

Holliday, V.T., ed. 1992. *Soils in Archaeology: Landscape Evolution and Human Occupation*, Washington, DC: Smithsonian Institution Press.

Howard, A.J. and Macklin, M.G. 1999. A generic geomorphological approach to archaeological interpretation and prospection in British river valleys: a guide for archaeologists investigating Holocene landscapes. *Antiquity* 73: 527–541.

Howard, A.J., Smith, D.N., Garton, D., Hillam, J. and Pearce, M. 2000. Middle to Late Holocene environments in the Middle to Lower Trent valley. In: Brown, A.G. and Quine, T.A., eds., *Fluvial Processes and Environmental Change*, Chichester: John Wiley & Sons, pp. 165–177.

Hudson-Edwards, K.A., Macklin, M.G., Finlayson, R. and Passmore, D.P. 1999. 2000 years of sediment-borne heavy metal storage in the Yorshire Ouse basin, northeast England. *Hydrological Processes* 13: 1087–1102.

Jackson, J., Norris, R. and Youngson, J. 1996. The structural evolution of active fault and fold systems in central Otago, New Zealand: evidence revealed by drainage patterns. *Journal of Structural Geology* 18: 217–234.

James, L.A. 1989. Sustained storage and transport of hydraulic gold mining sediment in the Bear River, California. *Annals of the Association of American Geographers* 79: 570–592.

James, L.A. 1999. Time and the persistence of alluvium: river engineering, fluvial geomorphology, and mining sediment in California. *Geomorphology* 31: 265–290.

James, L.A. 2006. Bed waves at the basin scale: implications to river management and restoration. *Earth Surface Processes and Landforms* 31: 1692–1706.

James, L.A. 2010. Sediment waves, bed waves, legacy sediment, and hydraulic myopia. *Geography Compass* 4/6: 576–598.

James, L.A. 2011. Contrasting geomorphic impacts of pre- and post-Columbian land-use changes in Anglo America. *Physical Geography* 32(5): 399–422.

James, L.A. 2013. Impacts of early agriculture and deforestation on geomorphic systems. In: Shroder, J.F., James, L.A., Harden, C. and Clague J., eds., *Treatise on Geomorphology*, vol. 13, San Diego, CA: Academic Press, pp. 48–67.

James, L.A. and Lecce, S. 2013. Impacts of land-use and land-cover change on river systems. In: Shroder, J.F. and Wohl, E., eds., *Treatise on Geomorphology*, vol. 9, San Diego, CA: Academic Press, pp. 768–793.

Jones, P.F., Salisbury, C.R., Fox, J.F. and Cummins, W.A. 1977. Quaternary terrace sediments of the middle Trent basin. *Mercian Geologist* 7: 223–229.

Keller, E.A. and Pinter, N. 1996. *Active Tectonics: Earthquakes, Uplift and Landscape*, Englewood Cliffs, NJ: Prentice Hall.

Kirkby, A. and Kirkby, M.J. 1976. Geomorphic processes and the surface survey of archaeological sites in semi-arid areas. In: Davidson, D.A. and Shackley, M.L., eds., *Geoarchaeology: Earth Science and the Past*, London: Duckworth, pp. 229–254.

Knight, D. and Howard, A.J. 1994. *Archaeology and Alluvium in the Trent Valley*, Nottingham: Trent and Peak Archaeological Trust.

Knighton, A.D. 1989. River adjustment to changes in sediment load: the effects of tin mining on the Ringarooma river, Tasmania, 1875–1984. *Earth Surface Processes and Landforms* 14: 333–359.

Knighton, A.D. 1991. Channel bed adjustment along mine-affected rivers of northeast Tasmania. *Geomorphology* 4: 205–219.

Knox, J.C. 1972. Valley alluviation in Southwestern Wisconsin. *Annals of the Association of American Geographers* 62: 401–410.

Knox, J.C. 1987. Historical valley floor sedimentation in the Upper Mississippi Valley. *Annals of the Association of American Geographers* 77 224–244.

Knox, J.C. 2006. Floodplain sedimentation in the upper Mississippi Valley: natural versus human accelerated. *Geomorphology* 79: 286–310.

Ladd, S.C., Marcus, W.A. and Cherry, S. 1998. Differences in trace metal concentrations among fluvial morphologic units and implications for sampling. *Environmental Geology* 36: 259–270.

Lang, A., Bork, H.-R., Mäckel, R., Preston, N., Wunderlich, J. and Dikau, R. 2003. Changes in sediment flux and storage within a fluvial system: some examples from the Rhine catchment. *Hydrologic Processes* 17: 3321–3334.

Lecce, S.A. and Pavlowsky, R.T. 2001. Use of mining-contaminated sediment tracers to investigate the timing and rates of historical floodplain sedimentation. *Geomorphology* 38: 85–108.

Leenaers, H., Schouten, C.J. and Rang, M.C. 1988. Variability of the metal content of flood deposits. *Environmental Geology and Water Sciences* 11: 95–106.

Lewin, J. 2010. Medieval environmental impacts and feedbacks: the lowland floodplains of England and Wales. *Geoarchaeology* 25: 267–311.

Lewin, J. and Macklin, M.G. 1987. Metal mining and floodplain sedimentation in Britain. In: Gardiner, V., ed., *International Geomorphology 1986, Part I*, New York: John Wiley & Sons, pp. 1009–1027.

Lewin, J., Davies, B.E. and Wolfenden, P.J. 1977. Interactions between channel change and historic mining sediments. In: Gregory, K.J., ed., *River Channel Changes*, New York: John Wiley & Sons, pp. 353–367.

Macklin, M.G. 1995. Archaeology and the river environment in Britain: a prospective review. In: Barham, A.J. and Macphail, R.I., eds., *Archaeological Sediments and Soils: Analysis, Interpretation and Management, 10th Anniversary Conference of the Association of Environmental Archaeology, 1989*, London: Institute of Archaeology, University College, pp. 205–220.

Macklin, M.G. and Lewin, J. 2008. Alluvial responses to the changing Earth system. *Earth Surface Processes and Landforms* 33: 1374–1395.

Macklin, M.G., Rumsby, B.T. and Newson, M.D. 1992. Historical floods and vertical accretion of fine-grained alluvium in the lower Tyne valley, Northeast England. In: Billi, P., Hey, R.D., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel-bed Rivers*, Chichester: John Wiley & Sons, pp. 564–580.

Magilligan, F.J. 1985. Historical floodplain sedimentation in the Galena River basin, Wisconsin and Illinois. *Annals of the Association of American Geographers* 75: 583–594.

Mantei, E.J., Ernst, R.L. and Zhou, Y. 1993. Comparison of metal homogeneity in grab, quartered, and crushed-sieved portions of stream sediments and metal content change resulting from crushing-sieving activity. *Environmental Geology* 22: 186–190.

Marcus, W.A. 1987. Copper dispersion in ephemeral stream sediments. *Earth Surface Processes and Landforms* 12: 217–228.

Meade, R.H., Yuzyk, T.R. and Day, T.J. 1990. Movement and storage of sediment in rivers of the United States and Canada. In: Wolman, M.G. and Riggs, H.C., eds., *Surface Water Hydrology. The Geology of North America*, vol. O-1, Boulder, CO: Geological Society of America, pp. 255–280.

Mordant, D. and Mordant, C. 1992. Noyen-sur-Seine: a mesolithic waterside settlement. In: Coles, B., ed., *The Wetland Revolution in Prehistory, Wetland Archaeological Research Project Occasional Paper 6*, Exeter: The Prehistoric Society and Wetland Archaeological Research Project, pp. 55–64.

Mossa, J. and James, L.A. 2013. Impacts of mining on geomorphic systems. In: Shroder, J.F., James, L.A., Harden, C. and Clague J., eds., *Treatise on Geomorphology*, vol. 13, San Diego, CA: Academic Press, pp. 74–95.

Ortloff, C.R. and Crouch, D.P. 1998. Hydraulic analysis of a self-cleaning drainage outlet at the Hellenistic city of Priene. *Journal of Archaeological Science* 25: 1211–1220.

Petit, F. 1995. Régime hydrologique et dynamique fluviale des rivières ardennaises. In: Demoulin, A., ed., *L'Ardenne, Essai de Géographie Physique. Livre en Hommage au Prof. A. Pissart*, Liège: Université de Liège, pp. 194–223.

Redman, C.L. 1999. *Human Impact on Ancient Environments*, Tucson, AZ: University of Arizona Press, 239pp.

Reece, D.E., Felkey, J.R. and Wai, C.M. 1978. Heavy metal pollution in the sediments of the Coeur d'Alene River, Idaho. *Environmental Geology* 2: 289–293.

Renfrew, A.C. 1976. Archaeology and the earth sciences. In: Davidson, D.A. and Shackley, M.L., eds., *Geoarachaeology: Earth Science and the Past*, London: Duckworth, pp. 1–5.

Rubey, W.W. 1938. *The Force Required to Move Particles on a Stream Bed, US Geological Survey Professional Paper 189-E*, Washington, DC: US Government Printing Office, pp. 121–140.

Rumsby, B.T. and Macklin, M.G. 1996. River response to the last neoglacial (the 'Little Ice Age') in northern, western and central Europe. In: Branson, J., Brown, A.G. and Gregory, K.J., eds., *Global Continental Changes: the Context of Palaeohydrology, Special Publication 115*, London: Geological Society, pp. 217–233.

Salisbury, C.R., Whitley, P.J., Litton, C.D. and Fox, J.L. 1984. Flandrian courses of the river Trent at Colwick, Nottingham. *Mercian Geologist* 9: 189–207.

Salvador, P-G. 1991. *Le Thème de la Métamorphose Fluviale dans les Plaines Alluviales du Rhône et de I'Isere Bassin de Malville et Ombilic de Moirans (Isère, France), Thèse Géographie et Aménagement*, Lyon: Université de Lyon III, 498pp.

Schiffer, M.B. 1987. *Formational Processes of the Archaeological Record*, Albuquerque, NM: University of New Mexico Press.

Sluse, P. 1996. *Évolution de la Rulles, de la Semois et de la Mellier au Cours des Cinq Derniers Siécles Grâce aux Résidus Métallurgiques de l'Industrie du Fer et par l'Etude des Cartes Anciennes, Mémoire de Licence en Sciences Géographiques*, Liège: Université de Liège, 206pp.

Sluse, P and Petit, F. 1998. Evaluation de la vitesse de déplacement de la charge de fond caillouteuse dans le lit de rivières ardennaises au cours des trois derniers siècles, à partir de l'études des scories métallurgiques. *Géographie Physiques et Quaternaire* 52: 373–380.

Thames Discovery Programme. 2012. http://www.thamesdiscovery.org/ (accessed 3 October 2015).

Thorndycraft, V.R, Pirrie, D. and Brown, A.G. 1999. Tracing the record of early alluvial tin mining on Dartmoor, UK. In: Pollard, A.M., ed., *Geoarchaeology: Exploitation, Environments, Resources, Special Publication 165*, London: Geological Society, pp. 91–102.

Thorndycraft, V.R., Pirrie, D. and Brown, A.G. 2004. Terrace and floodplain records of medieval and prehistoric tin mining on Dartmoor, SW England. *Geoarchaeology* 19: 219–236.

Thornes, J.B. and Gilman, A. 1983. Potential and actual erosion around archaeological sites in South-east Spain. *Catena Supplement* 4: 91–113.

Todd, M. 1973. *The Coritani*, London: Duckworth.

Tricart, J. and Vogt, H. 1967. Quelques aspects du transport des alluvions grossières et du façonnement des lits fluviaux. *Geografiska Annaler* 49: 350–366.

Vita-Finzi, C. 1969. *The Mediterranean Valleys: Geological Changes in Historical Times*, Cambridge: Cambridge University Press.

Vita-Finzi, C. 1988. *Recent Earth Movements: An Introduction to Neotectonics*, London: Academic Press.

Walter, R.C. and Merritts, D.J. 2008. Natural streams and the legacy of water-powered mills. *Science* 319: 299–304.

Waters, M.R. 1992. *Principles of Geoarchaeology*. Tucson, AZ: University of Arizona Press.

Waters, M.R. and Nordt, L.C. 1995. Late Quaternary floodplain history of the Brazos river in East-Central Texas. *Quaternary Research* 43: 311–319.

Wertz, J.B. 1949. Logarithmic pattern in river placer deposits. *Economic Geology* 44: 193–209.

Wolfenden, P.J. and Lewin, J. 1977. Distribution of metal pollutants in floodplain sediments. *Catena* 4: 309–317.

Wolfenden, P.J. and Lewin, J. 1978. Distribution of metal pollutants in active stream sediments. *Catena* 5: 67–78.

Zalasiewicz, J., Williams, M., Fortey, R., Smith, A., Barry, T.L., Coe, A.L., et al. 2011. Stratigraphy of the Anthropocene. *Philosophical Transactions of the Royal Society A* 369: 1036–1055

Zeuner, F.E. 1945. *The Pleistocene Period: Its Climate, Chronology and Faunal Succession, Royal Society Publications No. 130*, London: Roylal Society.

CHAPTER 4

Using historical data in fluvial geomorphology

Robert C. Grabowski[1] and Angela M. Gurnell[2]

[1]Cranfield Water Science Institute, Cranfield University, Cranfield, UK

[2]Queen Mary University of London, London, UK

4.1 Introduction

Long-term monitoring sites that document temporal changes in the landscape have rarely been established. The most notable example is the Vigil Network (Emmett and Hadley 1968; Osterkamp *et al.* 1991), an international network of areas including stream channels, hillslopes, reservoirs, precipitation and vegetation, on which periodic measurements are made and preserved. However, most studies of fluvial systems extend for periods of less than 5 years and, at best, provide detailed snapshots of the system or a small part of it. Therefore, for most studies, the only way to gain insights into the temporal variability of river channels in the longer term is by assembling and analysing historical data. Knowledge of previous conditions in a catchment, both along the river corridor and in the river channel, can provide valuable insights into contemporary channel behaviour. Historical analyses are required to establish channel and catchment conditions at one or more times in the past and to define times of major catchment, riparian and channel impacts, such as land use change, channelization and other engineering interventions. Consequently, many contemporary problems in fluvial geomorphology require a historical perspective, whether the concern is to understand natural patterns of channel form variation, to establish the nature of human impacts or to define benchmark conditions for channel restoration and management. For example, historical information can be useful in dating channel and catchment changes, documenting the nature and in some cases the rate of channel change, documenting changes in catchment conditions and human pressure on fluvial processes and forms, documenting channel response to and recovery from large floods and other disturbances, and so on. It is only in the context of an understanding of the channel's evolution that we can confidently interpret current conditions.

Useful information on a range of geomorphological questions can be provided by analysis of historical sources (Table 4.1). For example, Cooke and Reeves (1976) provided a useful early demonstration of the potential of historical analysis, by combining channel widths from early maps, using old buildings and other structures to determine erosion rates, using repeat stage-discharge rating records to demonstrate cross-sectional changes and reconstructing livestock densities from census reports, travel accounts and other sources to illuminate land-use changes. More recently, geomorphologists have benefited from advances in both historical geography (e.g. Hooke and Kain 1982), waterfront archaeology (e.g. Milne and Hobley 1981) and palaeohydrology (Gregory 1983; Gregory *et al.* 1987; Starkel *et al.* 1991), and through the latter's links with geoarchaeology (Goldberg and Macphail 2006) and palaeoecology (Berglund 1986). Useful overviews include a synthesis of information from a wide range of sources to examine the history of European rivers (Petts *et al.* 1989) and a critical review of the use of historical data sources for studying fluvial geomorphological change in the United States by Trimble (2012). Other useful papers include a review by Patrick *et al.* (1982) of methods for studying accelerated fluvial change, a review by Trimble (1998) of dating fluvial processes from historical data and artefacts, a review by Hooke (1997) of styles of channel change, a reconstruction by Large and Petts (1996) of a channel–floodplain system and demonstrations of the potential of historical analysis for river restoration by Sear *et al.* (1994) and Kondolf and Larson (1995).

This chapter addresses the range of sources that can be used to investigate changes in fluvial forms and processes over time; including documentary evidence, cartographic sources, topographic surveys and remotely sensed data. We present the data sources in increasing order of complexity, from individual human observations of a landscape or fluvial event up to scientifically derived two- and three-dimensional characterizations of catchment and river conditions. We start with documentary evidence (Section 4.2). Documentary sources encompass a diverse collection of records, from travel accounts and diaries to tax records and inventories of agricultural production, which in general give a one-dimensional snapshot of the catchment, river form or dimensions, or fluvial events (e.g. flood extent). Section 4.2 complements the evaluation of archaeological data presented in Chapter 3. In Section 4.3, the cartographic record is presented. Maps provide a wealth of two-dimensional information on river catchments and channels (e.g. land cover and channel position) and are the most commonly used data source for historical analysis. River topographic surveys are discussed in Section 4.4. Cross-sectional and long profile surveys allow

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

Table 4.1 Some examples of the use of different information sources for historical analysis.

Attribute	Source	Example
Channel planform		
Width	Maps	Gurnell *et al.* 1994
		Surian *et al.* 2009
Migration	Land surveys	Galatowitsch 1990
	Maps	Hudson and Kesel 2000
		Harmar and Clifford 2006
		Greco *et al.* 2007
Cut-offs	Botanical evidence	Everett 1968
	Surveys and travel accounts	Erskine 1992
	Maps	Hooke and Redmond 1989
Planform change	Maps and aerial photos	Comiti *et al.* 2011
		Hohensinner *et al.* 2013a,2013b
Channel depth and cross-section		
Long-profile/bedform	Navigation surveys	Large and Petts 1996
Channel incision	Topographic records	Piégay and Peiry 1997
		Rinaldi and Simon 1998
Cross-section	Bridge surveys	Kondolf and Swanson 1993
		Brooks and Brierley 1997
	Stage-discharge ratings and discharge measurement notes	Williams and Wolman 1984; Collins and Dunne 1989; Smelser and Schmidt 1998
	Repeat channel surveys	Petts and Pratts 1983
Land cover and riparian vegetation		
Land use	Land survey	Manies and Mladenoff 2000
	Land registry	Bender *et al.* 2005
Riparian vegetation	Cadastral surveys	Kondolf and Piégay 2007
	Maps	Hohensinner *et al.* 2004, 2011
		De Jager *et al.* 2013
	Travel accounts	Maser and Sedell 1994
	Ground photography	Beschta and Ripple 2006
Sediment		
Sedimentation	Datable artefacts and strata	Brown 2009
Sediment yield	Lake sediment chronologies	Davis 1976; Foster *et al.* 1985; Foster *et al.* 2011
	Reservoir storage changes	Trimble and Carey 1984
	Topographic records	Surian and Cisotto 2007
Inventory of engineering structures	Government records	Walter and Merritts 2008
	Maps and aerial photographs	Ziliani and Surian 2012
Climate and hydrology		
Climate	Diaries, log books and newspapers	Bradley and Jones 1992
Flood dates	Government documents	Uribelarrea *et al.* 2003
Hydrological conditions	Diaries, journals and newspapers	Snell 1938
	Water-level and flood records on buildings	Pfister 1992
Palaeofloods	Slack- *Downstream Effects of Dams on Alluvial Rivers* water deposits	Kochel and Baker 1988

the reconstruction of the three-dimensional form of the channel over time and are essential for investigations of base level changes. This section complements chapters on fluvial form (Chapter 11) and sediment budgets (Chapter 16). Finally, the emerging role of remotely sensed data in historical analysis is discussed in Section 4.5. Remote sensing is covered in detail in Chapter 6; however, it is becoming an important data source for historical analysis, so is briefly introduced here. In each section of this chapter, we give an overview of the sources of information, a brief description of how they are analysed, examples from the literature to illustrate their geomorphological applications and a general discussion of reliability and accuracy.

4.2 The documentary record

Documentary evidence serves two primary roles in the historical analysis of channel and catchment characteristics. The first is to extend analyses back in time prior to the collection of systematic survey data. In practice, this point in time ranges from the mid-19th to early 20th century when large-scale mapping programmes commenced in many countries, based upon topographic surveys that used accurate surveying, levelling and map-making techniques. The second is to corroborate observations or measurements from other sources (e.g. large-scale maps that pre-date modern mapping and surveying conventions) or to serve as a temporal or spatial benchmark with which to gauge river change (e.g. the location of a building in relation to historic channel position).

Historical information can be obtained from a variety of documentary sources. The specific sources available for a river depend on the country or region in which it is located, although typically these sources include land surveys, tax records, agricultural censuses, records from private estates and ecclesiastical sources, national government acts, legal proceedings and disputes over water rights, navigation passage and fisheries, written travel accounts and diaries, and photographs, drawings and landscape painting. There can be a degree of overlap between documentary and cartographic evidence as many surveys incorporated some element of mapping. This is especially true for land and tax (i.e. cadastral) surveys, which in many countries include both detailed reports on land ownership and utilization in addition to large-scale maps of land parcels.

Early documentary records in Britain

The nature of documentary evidence varies enormously, but the following British examples give a flavour of the variety and potential and also the quality of such records. Documentary evidence includes much early information related to property transactions, rents, tithes, etc., contracts and disputes and legal cases regarding the breach of these. The larger and wealthier a landowner, the more likely it was that documents would have been generated and the more likely it is that they will have survived. Thus, rivers within the largest, wealthiest estates can often have a large deposit of relevant documents including letters (both personal and business), property deeds and surveys with some maps and rentals. However, estate surveys were expensive and therefore were made only at extraordinary times, such as on change of ownership or in response to major agricultural improvements. Ecclesiastical documents can be valuable sources and often include information on fisheries, ferries, mill operations and floods. Large abbeys had major interests in economic farming practices and weather events affecting crops were of vital interest. Consequently, many abbeys produced important Chronicles, which record storms, floods and droughts from as early as the 13th century.

Thus, in Dugdale (1772, p. 150) there is evidence of the channelization of the Ancholme River and of the draining of its associated marshes (Fig. 4.1). In response to an enquiry from the king:

> Whereupon a jury being impanelled accordingly and sworn, did say upon their oaths, that it would not be to the damage of the said king, nor any other, but rather for the common benefit of the whole

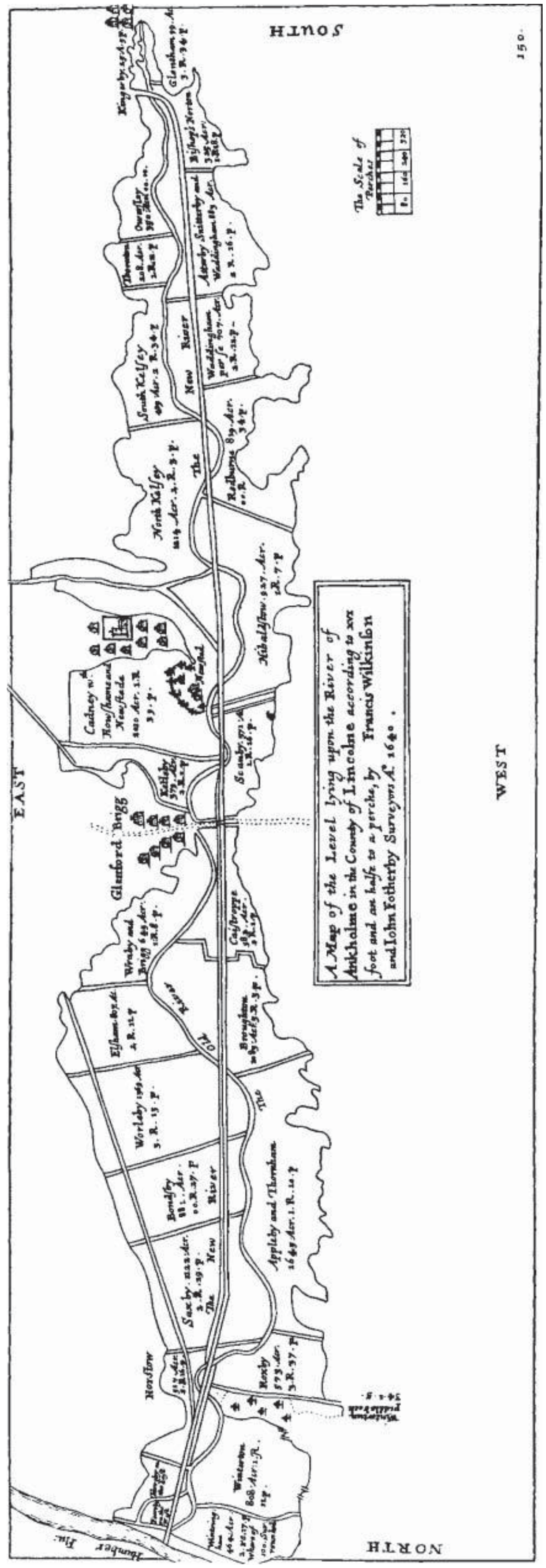


Figure 4.1 A survey of 1640 of the River Ankholme (now Ancholme) in eastern England showing an early channelization scheme and the natural course of the river. Source: Dugdale, 1772.

> county of Lincoln, if the course of that river, obstructed, in part, in divers places, from Bishop's Brigge to the river of Humber, were open. And they further said, that by this means, not only the meadows and pastures would be drained, but that ships and boats laden with corn and other things, might then more commodiously pass with corn and other things from the said river Humbre, into the parts of Lindsey, than they at that time could do, and as they had done formerly ...

The records include specific instructions: '... scouring the said channel from Glaunford brigge to the river of Humbre, to the breadth of XI feet, as it ought and want to be' (Dugdale 1772, p. 150).

A second example is provided by the reclamation of the wetland fens in eastern England, which once extended for 3400 km^2 (Butlin 1990). Camden (1586) records the general character of the area prior to drainage:

> All this country in the winter-time and sometimes for the greatest part of the year, laid underwater by the rivers Ouse, Grant, Nene, Welland, Glene and Witham ... it affords great quantities of turf and Sedge for firing; Reeds for thatching; Elders also and other water shrubs, especially willows, either growing wild or else set on the banks of rivers to prevent their overflowing ...

Of particular importance in England is the later Tithe Survey (ca 1830–1850), relating to the commutation of tithes previously paid in kind, as dues to support the local church. These comprise a large-scale map and a survey including the names of landholders, tenants and cottage holders, acreages, land use, fieldnames, parcel numbers and rental value. Other important documents of local administration were the Parliamentary Enclosure awards dating from the period 1750–1830, which combined an accurately surveyed map with a document of apportionment giving each landowner and tenant the parcels of land allocated to them. The Acts themselves are rather long, dense, legal documents but the associated correspondence can be informative, providing information on land use and value. Other useful Acts, together with associated correspondence, include Acts to improve the navigability of rivers, to build bridges and to build canals. The navigable rivers Acts of 1699 were particularly important for generating information on English rivers.

Often, documents generated during the formulation of such bills included detailed plans and surveys of the river supported by explanatory text. The historical maps that are available for Britain are listed by Hooke (1997, pp. 240–241). In addition, maps or charts for navigation date from 1795 when the office of Hydrographer was established at the Admiralty and the second Hydrographer, Captain Hurd, originated the *Charts of the Coasts and Harbours in all Parts of the World*. Surveys were also undertaken of the lower reaches of navigable rivers, and in England between 1810 and 1835 John Rennie published detailed channel plans as the basis for training rivers, including the Tyne, Ouse, Nene, Welland and Witham (Fig. 4.2) (Petts 1995, p. 7). All these historical sources provide information on the date and extent of channel modification, floodplain and wetland drainage and land-use change along the river corridor. Some early surveys, in addition to later ones, also present opportunities for quantitative analyses of channel planform and location.

The River Trent

Historical information available for the corridor of the River Trent provides an illustration of the range and types of documents that are often available, including local government, national government and ecclesiastical resources, in addition to family archives (Large and Petts 1996) (Table 4.2). The Trent was one of four 'royal' rivers. Rights of navigation were founded in a royal decree of Edward the Confessor of 1065 and disputes between navigators and mill and fishery interests ensured a long history of documentation. Legal cases in the Medieval and early modern period that were used to establish legal precedent have proved to be valuable resources, including documents concerning disputes over land between a number of large estates in the 15th century. A great deal of associated historical information survives, including the Harper–Crewe Papers relating to the Calke Abbey estate, for which rental details exist back to the 16th century, and the Every Papers, which contain some inscribed deeds and indentures relating to the period 1250–1600. Detailed surveys were also carried out at the time of extensive economic change and agricultural improvement in the 18th century.

The earliest known surveys of the river relate to the 1699 'Act for making and keeping the river Trent in the County of Leicester, Derby and Stafford navigable' – later known as the Paget Act. The Paget Act of 1699 provided for the making of a tow-path by which barges could be hauled, effectively changing the character of the riparian zone and requiring the maintenance of a morphologically smooth bank profile. The first detailed surveys of the river were carried out between 1761 and 1792 in order to develop the river for inland navigation. The surveys located and provided detailed low-flow depth soundings of 67 shoals along the river over a 90 km reach in the low-flow months of August and September (Fig. 4.3). The surveyor, William Jessop, made important observations on the fluvial geomorphology of the Trent (Petts 1995) and recommended works not only to self-scour the river but also to encourage overbank siltation, encouraging a natural process of channelization (Large and Petts 1996).

Other examples of the use of documentary sources in fluvial geomorphology

Documentary sources can be used to characterize historical land cover and land use at the catchment scale. For example, Bender *et al.* (2005) investigated changes in land use in southern Germany based on information from land registry records dating back to the early 19th century. They matched the land use records to parcels in cadastral maps to quantify changes in the areal coverage of land use types. However, for the United States, Trimble (2012) cautions against the use of land records or

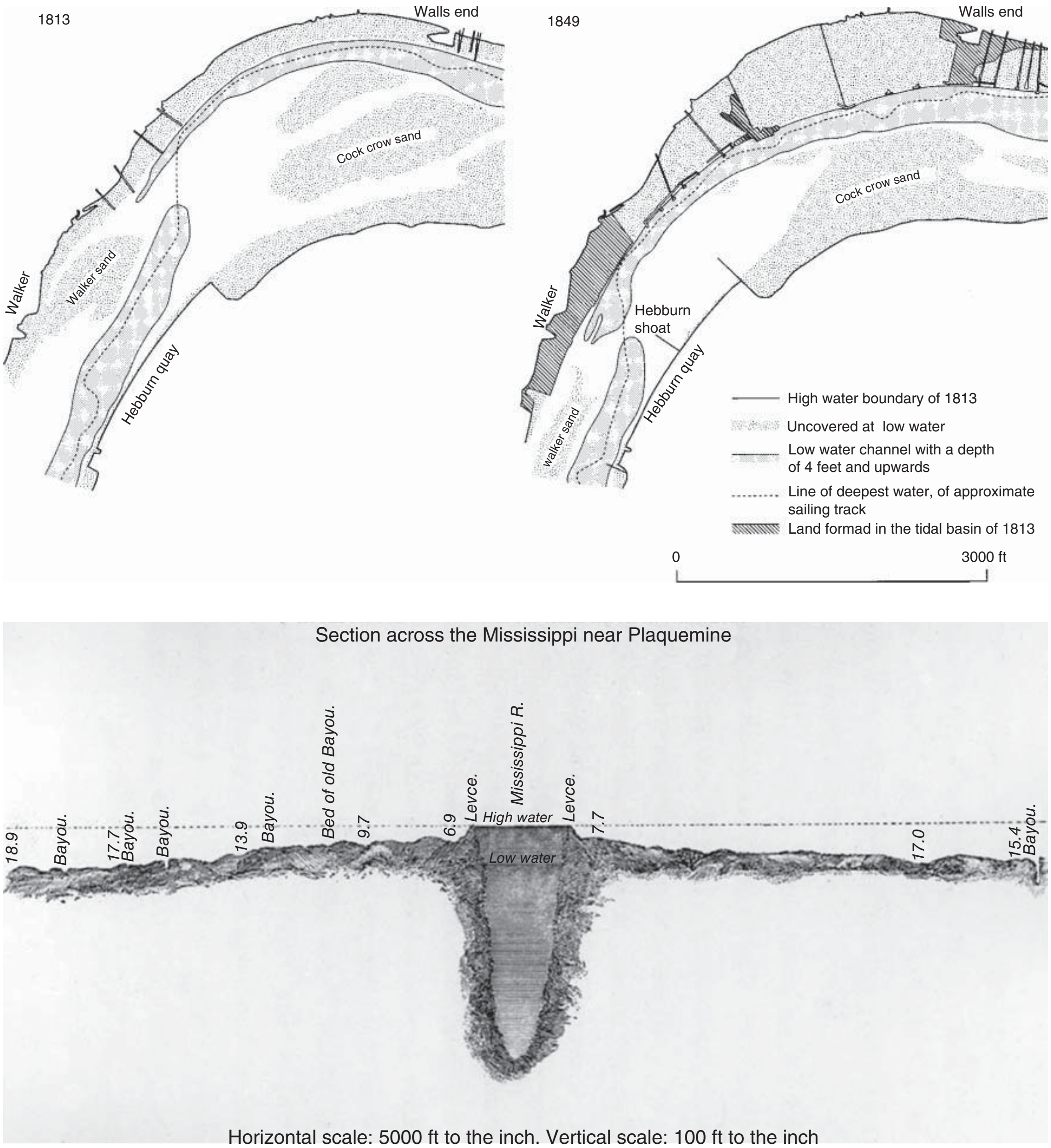


Figure 4.2 Examples of 19th century field-survey data. Above: surveys of the lower River Tyne in northeast England from Rennie's survey of 1813 and Calver's survey of 1849 (from Calver 1853). Below: cross-profile of the lower Mississippi (from Ellett 1853).

agricultural censuses to evaluate changes in land use over time. In his experience, records are often incomplete spatially (e.g. only farmed areas are included) and suffer from inconsistent terminology.

Historical evidence of riparian vegetation and large wood in rivers can come from early land survey documents and plans, travel accounts, cadastral surveys, ground photography and even landscape paintings and drawings. Travel accounts and early land surveys from the New World often describe vegetation and the presence of large wood in rivers that were minimally impacted by humans (Maser and Sedell 1994; Trimble 2008). Diaries of travellers in European countries also exist and could be a source of information on the use or condition of the river and riparian area (Hooke and Kain 1982). The investigation by Beschta and Ripple (2006) of changes in riparian vegetation and river planform in Yellowstone National Park (United States) following extirpation of wolves is an excellent example of the complementary use of historical ground photography with remote sensing data (in this instance aerial photography) to resolve vegetation dynamics.

Historical records can be used to investigate the magnitude and frequency of flood events. Despite some notable exceptions,

Table 4.2 Documentary sources available for the River Trent corridor.

1200–*Deeds and private legal papers; ecclesiastical sources*	
	Monastic Chronicles (1200–1713; incomplete record)
	Every Papers (1620–1890)
	Personal correspondence (1630–1890)
	Tithe surveys (1830–1850)
1699–*National Government Acts*	
	To improve navigation (1699, 1740, 1781, 1783)
	To build bridges (1758, 1835)
	To build canals (1766, 1777, 1793)
1750–*Local Government surveys*	
	Enclosure surveys (1750–1830)
	Topographical reports (1800–1820)

discharge records generally exist only from the late 19th century, so other documentary sources need to be assessed to extend the analysis further back in time. For example, Uribelarrea *et al.* (2003) gathered water stage data associated with historical flooding from a range of documentary sources as part of a study of channel change in two rivers in central Spain. Information mentioned in these documentary sources included sites or landmarks reached by a flood, areas of the floodplain that experienced flooding, areas or landmarks that were not flooded and estimates of flood severity in comparison with earlier floods. Discharges were then estimated from the historic flood levels using a one-dimensional hydraulic model and integrated into the gauging station records to produce a timeline of flooding dating back to 1557.

Historical inventories of water storage and diversion structures, sediment control structures and sediment-related activities within a catchment can be created to support investigations of temporal change (e.g. Boix-Fayos *et al.* 2007; Ziliani and Surian 2012). Large dams and water diversion schemes can significantly alter flow regimes and the magnitude, frequency and timing of peak flows. For example, coarse sediment delivery and transport are impacted by dams, check dams, weirs and torrent controls, whereas fine sediment may be influenced additionally by drainage ditches and artificial levees. Depending on the catchment history, it may also be pertinent to acquire data for sediment-related activities within the channel, such as records detailing the quantity and location of sediment dredging or mining from the channel (e.g. Wishart *et al.* 2008; Martin-Vide *et al.* 2010). One method of summarizing these anthropogenic pressures is to develop a catchment chronology (recent examples include Ziliani and Surian 2012 and Downs *et al.* 2013) (Fig. 4.4). Typically, a chronology identifies the type of anthropogenic activities that are present in the catchment, when they began and ended, and whether the level of activity intensified or diminished over time. The chronology can integrate information from all sources in the historical analysis (e.g. land cover, riparian vegetation, channel dimensions, river discharge, major flood or drought events) and serve as tool to identify the underlying causes of temporal change in the river channel and floodplain.

An excellent example of how a historical inventory of flow control structures can be used in fluvial geomorphology comes from the work of Walter and Merritts (2008) on mid-Atlantic streams in the United States. They used county records, historical maps and historical photographs to determine the number and location of mill dams, and also evidence from later mill acts to document the negative impact that the high density of dams was having on river flows and sedimentation at the time. This historical work provided the context for further topographic, stratigraphic and sedimentological research, which concluded that mill dams and the conversion of forests to agriculture led to massive sedimentation in the floodplain that dramatically altered the channel and floodplain morphology. The study concluded that the current single-thread, deeply incised channels common to this area had most likely been anabranching rivers flowing through forested wetlands prior to settlement. This hypothesis changes the underlying assumptions of contemporary channel processes in these systems and has implications for river restoration.

Problems of data reliability and accuracy

The reliability of data extracted from documentary sources varies depending on the original purpose of the data source and also its age. In general, accuracy increases with decreasing age of the source and for time periods preceding systematic

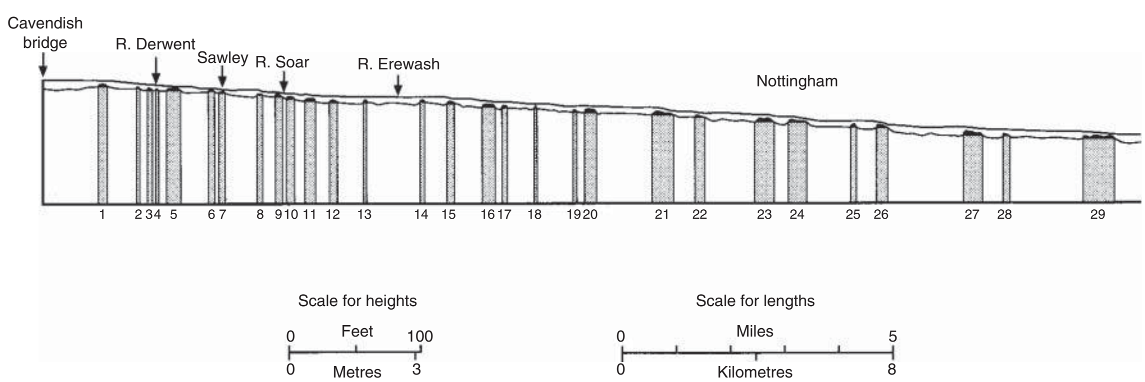


Figure 4.3 A section of William Jessop's survey of the River Trent showing the location and spacing of 29 shoals in 1792. Source: Jessop, 1782.

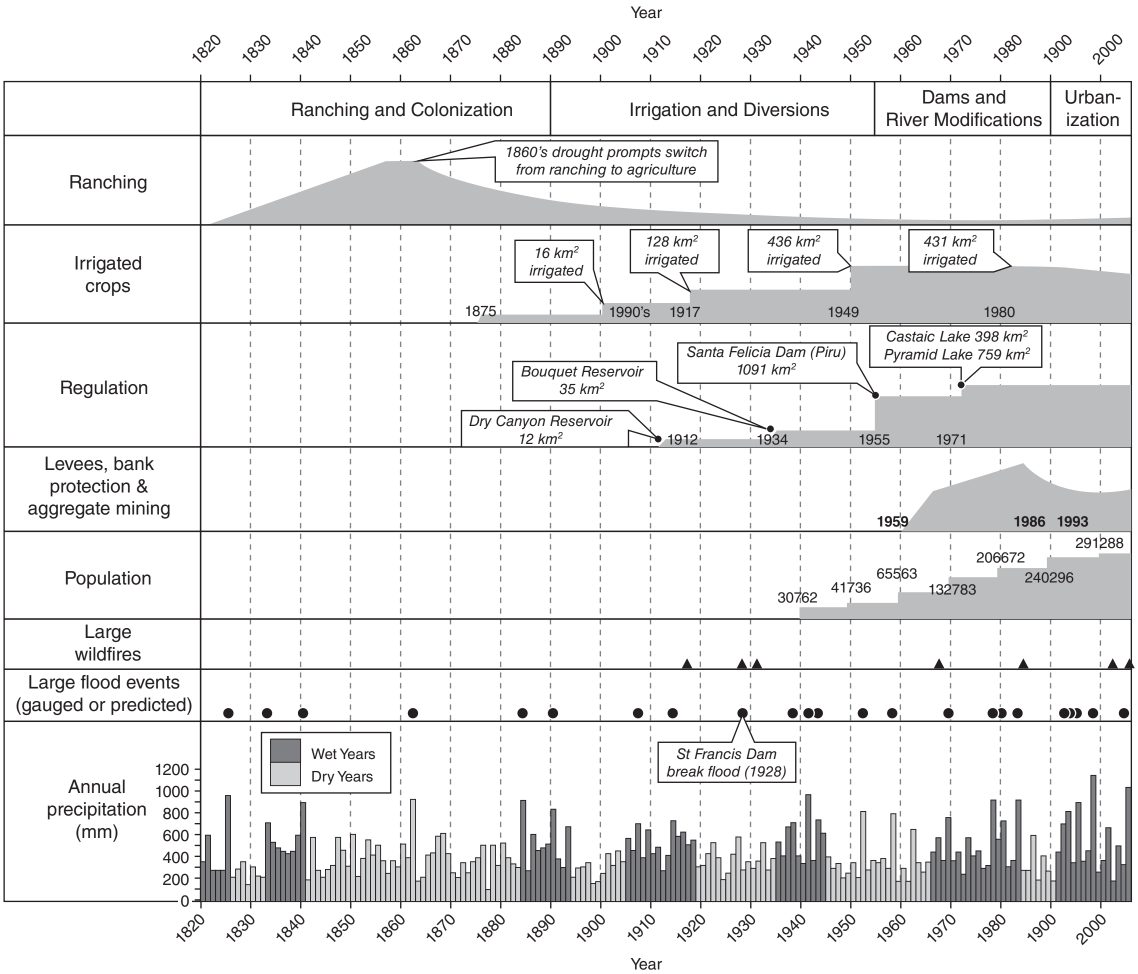


Figure 4.4 Chronology of anthropogenic pressures and natural disturbances on the Lower Santa Clara River, USA. Source: Downs *et al*, 2013. Reproduced with permission of Elsevier.

survey techniques (approximately mid- to late 19th century), the availability and consistency of information are generally poor over both space and time.

There are three key aspects concerning data availability and consistency to consider. First, detailed information is often available only for parts of catchments or some reaches of a river or for some rivers, and there are questions about the representativeness of the information that is available. Such information often focuses upon river reaches experiencing particular problems relating to flood control, land drainage or navigation. Second, there are problems caused by changes over time in survey or recording conventions, such as the areal units for which data were collected and the measurement technique or recording procedures employed in different surveys. Experience in reconstructing climate records (Bradley and Jones 1992) shows that problems can also arise because of changing calendar conventions. Furthermore, discrepancies in descriptions and records can reflect changing perspectives of surveyors, including differing cultural attitudes towards natural resources (Hooke and Kain 1982).

Clearly, all sources require careful scrutiny and verification. Harley (1982) considered that the scholarly evaluation of historical evidence must involve reference to the context of that evidence – why and for what purpose was it collected? Topographic and map survey data are discussed later and so discussion here is restricted to the use of qualitative data sources. Of prime importance is data verification. Often documents contain a mixture of both valuable and worthless information. The latter includes inaccurate or uncertain dating of events or distortions or amplifications of original observations. Only if the observations are faithful in both time and space are they likely to be reliable and valuable for geomorphological interpretation. However, even if information can be verified, Bradley and Jones (1992, p. 6) emphasize the difficulty of ascertaining exactly what the information means. Terms such as ‘flood’, ‘frozen river’, ‘drought’ and ‘summer’ or ‘winter’ used in the past may not be equivalent to terms employed in modern-day observations. In some cases, for example, the term 'winter' has been used for the period of snow cover rather than for specific months. Qualifying words such as 'unprecedented', 'extreme', 'in living memory', 'extensive' or 'deep' can be ambiguous. Bradley and Jones (1992) suggest that a solution to this problem is to use content analysis to help isolate the most pertinent and unequivocal aspects of the historical source. Content analysis provides an objective approach to assessing the frequency of use of descriptive terms and the use made of qualifying terms (for an application to climate data, see Pfister 1992).

4.3 The cartographic record

In this section, we focus exclusively on maps and the planimetric data that they can provide for historical fluvial geomorphological analyses. Other plan sources, most importantly aerial photographs and multi- and hyperspectral data, are discussed briefly in Section 4.5. Remotely sensed data sets are presented in more detail in Chapter 6.

The cartographic record contains an abundance of maps created for an array of different purposes. Maps were produced for many reasons, including exploration and land surveying prior to settlement (e.g. land survey township plats in the United States), military and defence purposes, as part of the management of estates, for land and tax registries, to record land use and agricultural output, and as part of national mapping programmes. Their role in the historical analysis of fluvial geomorphology depends on the original purpose of the map, the survey methods and mapping conventions employed, the map scale and the map accuracy (positional, attribute and temporal). In ideal circumstances, maps can be used for quantitative analysis of temporal changes in channel position, planform characteristics and dynamics and catchment land cover and riparian vegetation. However, when inaccuracies and uncertainties associated with historical maps exceed the magnitude of change being detected, maps can provide only a qualitative assessment of channel form or features.

Historical analysis of channel planform is typically conducted by overlaying maps of different ages within a GIS (i.e. diachronic analysis). The first step in this process is to conduct an internal and, if possible, an external check of map accuracy (Hooke and Kain 1982, Chapter 3). If the map passes these checks, it is scanned into a digital format at a resolution sufficient to ensure that information is not lost in the transfer (e.g. colour maps, minimum 400 DPI and 24-bit colour; Library of Congress 2006). Next, the map is imported into a GIS and registered to a map projection. Map registration is typically accomplished by matching landmarks on the historical map with those on a modern large-scale map, a process known as georeferencing (Hu 2010). Geometric transformations are used in this process and can involve alterations to the scale, horizontal displacement and rotation of the historical map (Manzano-Agugliaro *et al.* 2013). When georeferencing, the choice of landmarks is important; river banks and other features that could have moved are not suitable, whereas bridges and buildings generally make good georeferencing points. Some difficulty may arise when working with very old maps or in remote areas where few, if any, shared landmarks exist. A possible solution is to work back in time from the current large-scale maps in a regressive–iterative approach that bases the georeferencing on landmarks in the next most recent map (Hohensinner *et al.* 2013b). The spatial distribution of georeferencing points, the precision of those locations (e.g. a building centroid versus the northwest corner of the building) and the types of transformation used will influence the positional accuracy of the map and thus uncertainties in features measured from it. In general, georeferencing points should be spaced evenly over a map to minimize distortion (Hu 2010) and a linear transformation should be used unless there is evidence of significant warping in the original map document (Gurnell *et al.* 1994). Finally, the diachronic analysis can be conducted. This typically involves digitizing the boundaries of features from the map (e.g. river banklines) and comparing their positions with earlier ones or quantifying an attribute of the features (e.g. channel width, land cover, sinuosity index) and investigating how the attribute changes over time.

Some examples of using maps to study channel change

Maps are commonly used in studies of channel and catchment change and the following examples give an indication of the types of information that can be extracted.

Temporal changes in planform morphology and channel width have been identified from maps in numerous studies. For example, many historical studies of braided rivers in Europe have used maps to assess the impacts of gravel mining and sediment control structures on the width of the active braidplain, and in some cases have recorded a shift from braiding to wandering planforms (e.g. Surian 1999; Winterbottom 2000; Wishart *et al.* 2008; Piégay *et al.* 2009; Comiti *et al.* 2011; Ziliani and Surian 2012). These studies often use maps from a variety of sources that have different scales and positional accuracies. For example, studies on Italian rivers have used a combination of 19th century military maps, regional maps and aerial photographs to describe planform changes over the last 200 years (Surian 1999; Surian *et al.* 2009; Comiti *et al.* 2011; Ziliani and Surian 2012). Although positional errors in these maps have been estimated to be on the order of tens of metres [root mean square error (RMSE) = 15–20 m], these are small relative to the change detected in the highly dynamic braided rivers that were studied (Comiti *et al.* 2011).

Diachronic analysis of river position can be used to investigate other aspects of planform dynamics, including to quantify the area that has been eroded or deposited by the river over a time period (Gurnell *et al.* 1994; Kummu *et al.* 2008), to investigate meander migration and cut-off dynamics (Hooke and Redmond 1989; Hudson and Kesel 2000; Harmar and Clifford 2006; Hooke 2007) (Fig. 4.5) or to estimate the erodible corridor of the river to

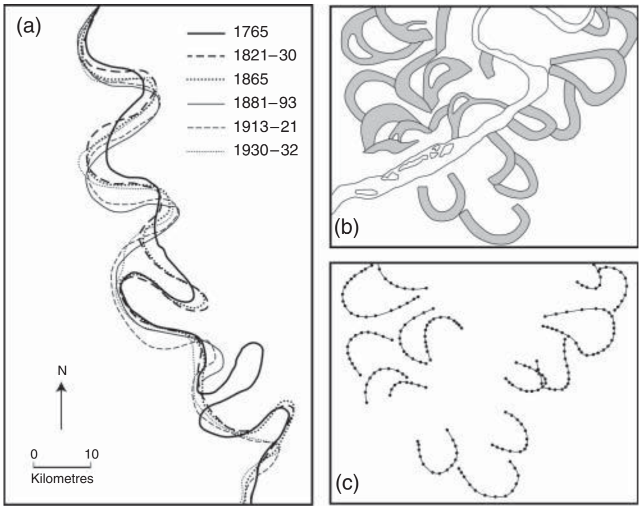


Figure 4.5 Historical maps of (a) river planform and (b, c) meander cut-offs of the Lower Mississippi River from earlier studies. Source: Harmar and Clifford, 2006. Reproduced with permission of Wiley.

give an indication of the likelihood of lateral movement (Piégay *et al.* 2005). A recent and comprehensive example of historical analysis using maps comes from the Danube River in Vienna, Austria, where Hohensinner *et al.* (2013a,b) used the rich cartographic record available for the city and surrounding landscape to reconstruct the morphology of the river and floodplain back to 1529. This detailed work revealed that, prior to river regulation and urbanization, the Danube was an anabranching river that flowed through forested wetlands. Moreover, it was possible to quantify changes in the areal coverage of river channels over this period, and also changes in channel and floodplain geomorphic features (e.g. gravel bars and backwaters).

Land cover and land use derived from topographical surveys, agricultural censuses and cadastral surveys and maps can provide important information on riparian and floodplain vegetation cover. For example, research by Hohensinner *et al.* (2004, 2011) on an upstream section of the Danube used detailed topographical maps from a variety of sources (Austrian federal, provincial, aristocracy, monastery and private archives) to investigate changes in the areal coverage of riparian and aquatic habitats over time. This information was combined with site-specific ecological data to estimate changes in habitat type, age and turnover (i.e. succession). In the United States, Greco *et al.* (2007) used a variety of historical maps and aerial photographs dating back to 1870 to estimate the age of floodplains created by the meandering Sacramento River, and De Jager *et al.* (2013) used historical land cover maps and aerial photographs to investigate land cover changes in the floodplain of the upper Mississippi River since 1890.

When channel planform characteristics and vegetation are examined simultaneously, it allows researchers not only to detect changes in these attributes but also to identify potential causal linkages. For example, Kondolf and Piégay (2007) used 19th century cadastral maps, 20th century topographical maps and recent aerial photography to characterize the riparian vegetation cover along the Eygues River in France. From these sources, they extracted the location of the channel, channel widths, the presence of vegetation and, in some instances, the type of vegetation from descriptive land use terms (e.g. *oseraie* or *hermes* described willow stands). By developing a chronology of river and riparian land-use change, the study was able to conclude that a reduction in local population and associated agricultural and grazing pressure decreased the sediment supply, which in turn resulted in channel incision and narrowing. The establishment of riparian vegetation followed the planform changes.

General issues of accuracy

Maps are simply abstractions or generalizations of reality that have been produced with a specific purpose in mind. Therefore, it is important to avoid attempts to extract more information from a map than was there in the first place. National and regional map-making agencies usually provide detailed manuals on the survey and mapping conventions used in map production and frequently give estimates of the accuracy of their products. These sources should be used to assess whether the purpose of a particular analysis can be met by the information provided in specific maps. For example, only water courses that are 5 m or more wide are shown to scale with two lines marking their banks on UK Ordnance Survey 1 : 10,000 scale maps, whereas the threshold is 1 m on 1 : 1250 scale maps (Harley 1975). A careful consideration of the accuracy and conventions built into map production can provide the basis for the extraction of quantitative information from the most unlikely sources. For example, Gurnell (1996) describes the use of a cover-abundance scale to extract quantitative spatial information from spatially distorted River Corridor Survey maps, which are essentially sketch maps produced for the UK Environment Agency to describe the biogeomorphological characteristics of 500 m stretches of river course. However, even if the intentional limitations of maps are taken into account, a variety of other errors can be introduced inadvertently at various stages in map production, which may have importance for geomorphological interpretation.

There are three fundamental dimensions of spatial data: space, attribute and time (Chrisman 1991) or 'where something was observed', 'what was observed' and 'when was it observed' (Flowerdew 1991) . The following account indicates some of the intentional and inadvertent errors associated with all of these three dimensions, which accumulate into a total map error.

Positional accuracy

The first constraint on positional accuracy is the technical limitations of the surveying equipment and the methodology employed at the time of the original field survey. However, perhaps more important are the conceptual errors that may have been introduced during field survey, air photograph interpretation or the interpretation of information from other sources. The surveyor frequently has to make decisions about the location of features or boundaries. Such decisions are particularly difficult in relation to natural features, which rarely have crisp boundaries. Some features, such as agricultural fields, may have clearly defined boundaries. However, other features, such as soils and vegetation, often grade gradually from one type to the next across transition zones, but the surveyor is still required to map a boundary. Even natural features with apparently crisp boundaries are usually 'fuzzy' in practice. For example, it may be straightforward to identify the position of a river bank where the bank is vertical, but difficulties can arise where the river bank consists of a gently sloping aquatic–terrestrial transition or a sequence of benches, slumps and terraces. As a result, conventions are usually devised to define boundaries. For example, river channel boundaries are defined by the UK Ordnance Survey in relation to the 'normal winter level' (Harley 1975), but there is still great potential for error in applying such conventions. The timing of the field survey is likely to be an important influence on the accuracy with which the 'normal winter level' is determined. 'If, therefore, the stream is surveyed in summer it is the permanent channel, eroded of vegetation, rather than the water width, which is measured' (Harley 1975, p. 44). Similarly, in semi-arid regions where channels are strongly influenced by infrequent, larger floods, definition of the unvegetated, active channel may vary in width depending upon the time of the survey in relation to the time of the last large flood.

Once the information for the map has been assembled, there is a range of error sources associated with translating the surveyed information into a map. All maps have a spatial reference system, based on a map projection, which translates latitude and longitude on the curved surface of the Earth onto a flat map sheet. Thus maps for the same area and at the same scale but based on different map projections are not directly comparable and, indeed, may vary in their spatial scale from one part of the map to another. Similar, but more severe, problems arise when using information derived from photographs. If the photographs are oblique, projection problems arise as a result of varying spatial scale over the photographic image. Even with vertical photographs, significant distortions occur with increasing distance from the centre of the image and with differences in altitude of the terrain.

Another source of positional error relates to the map scale. Scale determines the smallest area that can be drawn and recognized on a map. It is not possible to locate any object more accurately than the width of one line on the map. This determines the resolution of the map, which, assuming a minimum line width of 0.5 mm, is 5, 25 and 50 m for map scales of 1 : 10,000, 1 : 50,000 and 1 : 100,000, respectively (Fisher 1991). Clearly, this places a limit on the accuracy with which locations can be measured from a map, but there are many other factors that further degrade the locational accuracy of the map. For example, the map scale also influences whether or not features are shown on a map. Thus maps of soil, vegetation or rock types, which may be extremely variable over small areas, have to be based on a minimum mapping unit – the smallest area that can be represented on the map. Features smaller than the minimum mapping unit must either be merged with adjacent areas so that the map reflects dominant classes or if they are particularly important to the map theme, they can be represented by symbols or can be spatially exaggerated so that they can be mapped. This leads to the issue of information generalization, which is used to ensure the visual clarity of a particular map. For example, information may be omitted or spatially smoothed, even when it relates to areas significantly greater than the minimum mapping unit if inclusion of the information is detrimental to map clarity. As a result, not only do different types of thematic map at the same spatial scale represent the same information to different levels of detail, but also different editions of the same thematic map may present very different quantities of information on the same features. For example, Gardiner (1975) showed that the length of streams depicted on 1 : 25,000 scale UK Ordnance Survey topographic maps varied greatly with the map edition. The stream length ratio between the Second Series and the Provisional Edition varied between 1.10 and 1.80 for a sample of map sheets from different areas of Great Britain. Furthermore, several papers (Ovenden and Gregory 1980; Burt and Gardiner 1982; Burt and Oldman 1986) have explored the accuracy with which headwater stream networks are depicted on Ordnance Survey 1 : 10,560 and 1 : 10,000 scale maps. These papers illustrate that extreme care must be taken in interpreting such information from different map editions and for different geographical locations.

A final point relates to the boundaries of map sheets. Traditional map series were often designed as a series of individual map sheets with no guarantee of conformity across the margins of the maps. This can lead to many anomalies on map sheet margins, which simply reflect decisions relating to the generalization and presentation of features on the individual sheets. All of these factors illustrate that although the resolution of a map is fundamentally dictated by map scale, there are a range of other factors that vary within and between maps and that influence the positional accuracy of the features that are depicted.

Although most of the comments made above relate to printed maps, it is important to remember that digital map products are subject to the same surveyor errors if human survey is needed to derive the data (e.g. geological or soil maps). Furthermore, these products are frequently derived, at least in part, from paper maps and so incorporate all of the potential errors discussed above, with the addition of digitizing error. In addition, many digital products are provided in a grid format, which in many cases has been interpolated from non-gridded data derived from printed maps, so interpolation error is yet another addition to the list of possible error sources.

Attribute accuracy

The accuracy of mapped attributes varies according to the measurement scale employed. If the attribute is measured on a continuous scale (e.g. precipitation), it can only be recorded on the map to a given level of accuracy. Particular problems arise for features, such as elevation, which occur everywhere and which are often represented on maps by isolines. The first problem relates to the precision of the attribute estimates on the mapped isolines. For example, the technical specification of the contours and grid data of the UK Ordnance Survey OS Terrain® digital product is given as RMSE less than 2 m. Even if the attribute values along the isolines are completely reliable, values of the attribute and error margins for points on the map that are located between isolines are difficult to assess. If the attribute is categorical, exact recording is possible. However, as in the case of soil maps, mapped categories are frequently based on a classification, which may not represent the level of discrimination required by the user and which is also open to inaccurate interpretation by the surveyor.

Temporal accuracy

Every map relates to a particular survey date and so is always out of date by the time it is published. Because surveys are undertaken in different places at different times, the extent to which any particular map is out of date varies between different map sheets, even within the same thematic map series. Whereas these sources of temporal inaccuracy can be determined from information provided with the map, other sources of temporal (in)accuracy are more difficult to detect. Many maps are declared to be partial revisions of their predecessors or, more seriously, Carr (1962) provides examples of the use of information from previous maps, without acknowledgement. In both of these cases, even assuming that the partial resurvey is accurate, there is no guarantee that the information depicted on the map is from the indicated date of survey.

A further time-related source of error in paper maps results from shrinkage and distortion of the paper over time and distortion resulting from the use of copies of the original map.

Assessing accuracy

The above discussion illustrates that it is important to devote some consideration to map accuracy if spurious geomorphological conclusions are to be avoided.

Positional accuracy of a historical map can be assessed by comparing positions on the map with actual locations on the ground or with their location on a more recent map or digital product with higher accuracy. This generates a series of displacements, which can be analysed for both systematic and random or residual error components (Chrisman 1991; Mount *et al.* 2003). The former can often be removed by geometrical transformation in a GIS, whereas the latter can then be quantified to provide 'error margins' for positional information extracted from the map. This type of approach is used by mapping agencies to check the accuracy of their products.

RMSE is the most commonly used metric for representing positional error. A single RMSE value is estimated for a map during georeferencing and is often used as a minimum threshold for change detection (e.g. Winterbottom 2000). However, this estimate of error does not incorporate all of the possible sources of error and may underestimate the uncertainty in measurements derived from the maps. Errors associated with the original field survey, cartographic representation and digitization can be combined into a total error estimate using the quadratic sum (Cheung and Shi 2004; James *et al.* 2012). Furthermore, errors propagate when features are compared over time to detect change, so errors should be summed to yield a minimum threshold of change, which can be divided by time to yield a minimum rate of change (Del Rio and Javier Gracia 2013). Other approaches to estimating positional error include that proposed by Mount and Louis (2005), which allows for anisotropy of the random error component and the spatially explicit approaches to accuracy and feature change detection in historical maps proposed by Tucci and Giordano (2011) and Manzano-Agugliaro *et al.* (2013).

As discussed above, additional uncertainty arises when the features of interest are ill-defined and so do not have sharp boundaries. In this case, 'ground-truth' information may be required to estimate appropriate additional error margins relating to positional uncertainty.

Attribute accuracy can be tested in a similar manner to positional accuracy when the attribute is continuous (e.g. elevation). Where the attribute is categorical, the construction of a mis-classification matrix based upon map and 'ground-truth' information for the same sites can help to assign percentage errors to different attribute classes.

4.4 The topographic record

Following on from the plan view of rivers and catchments provided by historical maps, this section discusses the historical topographic data that help to build a three-dimensional view of channel form and change over time. Detailed topographic surveys of rivers began approximately 100–150 years ago in response to their development for navigation, water resource use and flood control. These surveys have produced two kinds of data that are of particular use in the geomorphological study of river channels: cross-sectional and long profile surveys.

A cross-section is a two-dimensional representation of the channel form oriented perpendicular to the flow direction. Cross-sections provide the basis from which morphometric indices (e.g. width, depth, thalweg position, water surface level and bed altitudes, channel asymmetry) or hydraulic indices (e.g. bankfull cross-sectional area and hydraulic radius) can be calculated (e.g. Gurnell 1997a). In some rivers, a network of cross-sections was established in the past and has been monitored regularly over long periods to assess changes in bed levels or channel widths (e.g. Brenta River, Italy; Surian and Cisotto 2007). In others, cross-sections may be limited to infrastructure that spans the river (e.g. bridges or dams) or associated with specific local management problems (Kondolf and Swanson 1993; Brooks and Brierley 1997; Erskine 1999). Access to cross-section data may be gained via national scientific agencies (e.g. United States Geological Survey), national transportation authorities or local government offices that hold planning documents associated with infrastructure construction.

Long profiles are two-dimensional representations of channel form oriented parallel to the flow direction and are intended for the study of slopes (channel bed slope, slope of the energy line for a given discharge). They may be directly constructed from a longitudinal topographic survey or may be derived from cross-section surveys that have been regularly distributed along the river channel. In both cases, the horizontal distance (x-axis) by which every point of altitude is referenced is the distance along the channel centre line derived from direct measurements in the field or from estimates from large-scale maps. In contrast, the values of altitude presented in the long profiles (y-axis) may vary: (i) altitudes almost always represent the water surface when the long profile has been surveyed along the river from upstream to downstream; (ii) when data are derived from cross-sections, the altitude may represent the water level, the average level of the bed or the elevation of the thalweg (Fig. 4.6). The water surface level is strongly dependent on the hydrological regime and hydrometeorological events. For reasons of convenience, historical topographic surveys were generally made at low flows, unless flood levels were the focus of the study, as for example when the survey was to be used to calibrate a hydraulic model. The average bed level is an altitude that smooths out the shape variability or asymmetry of the channel. It is frequently used when cross-sections are available to underpin estimation of the average bed level. The thalweg altitude or altitude of the lowest point of a channel cross-section provides useful historical information on the position of pool-riffles and their changes.

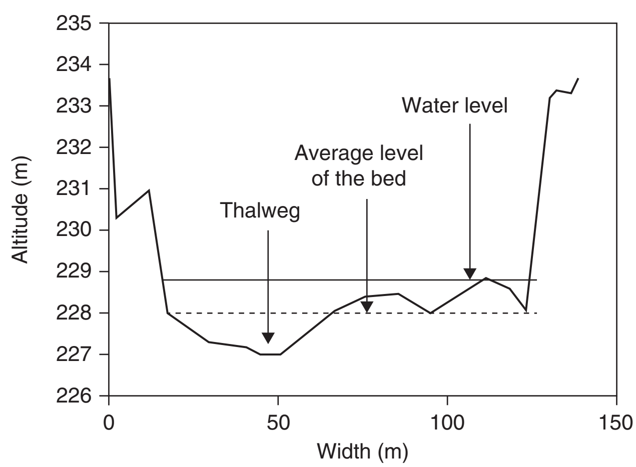


Figure 4.6 Data surveyed or calculated on a cross-section.

Some examples of using topographic records to study channel change

A common use of historical topographic records is to describe channel aggradation or incision linked to hydrological or sediment load changes.

Initially, long profile comparisons were used for studying complex readjustment of channel morphology below reservoirs (Petts 1979; Williams and Wolman 1984). More recently, fluvial geomorphologists have explored historical topographic data from archives to derive indices of natural or anthropogenic river metamorphosis. For example, Bravard (1987, 1994) demonstrated aggradation of the upper Rhône River over the 19th and 20th centuries in association with the downstream extension of braiding associated with the reworking and delivery of sediment from former glaciated basins. In the French Alps between 1840 and 1950, the longitudinal embankment of most rivers at a time of abundant bedload supply, associated with the climatic degradation of the Little Ice Age, frequently led to channel aggradation (Gemaehling and Chabert 1962). On the Isère River close to the city of Grenoble, this phenomenon has been particularly well documented by civil engineers. Topographic records allow the channel aggradation to be quantified at 1–2 cm per year between 1880 and 1950 (Blanic and Verdet 1975). In the last 20 years, topographic records have mainly been used to document channel incision and its spatial distribution. Such records have been particularly effective in documenting rapid, deep incision, which, for example, has reached up to 2 m, mainly as a result of the impact of gravel mining (Peiry 1987; Kondolf and Larson 1995; Rinaldi and Simon 1998; Rinaldi 2003; Surian and Cisotto 2007; Ziliani and Surian 2012).

From a technical point of view, a classical way to undertake a diachronic analysis of geomorphological changes is to superimpose cross-sections or long profiles on the same graph (e.g. Fig. 4.7). When differences in altitude between two long profiles are moderate, it is often more effective to graph positive and negative deviations in altitude. These differences in altitude can be extrapolated from long profiles systematically, at a constant horizontal interval (e.g. every 250 m to 1 km according to the

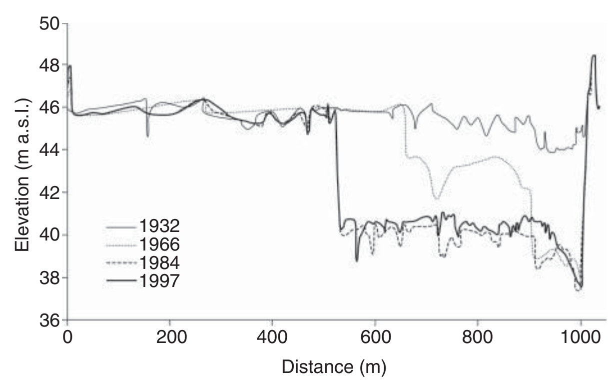


Figure 4.7 Change in cross section form and bed level in one reach of the Brenta River from 1932 to 1997. Source: Surian and Cisotto, 2007. Reproduced with permission of Wiley.

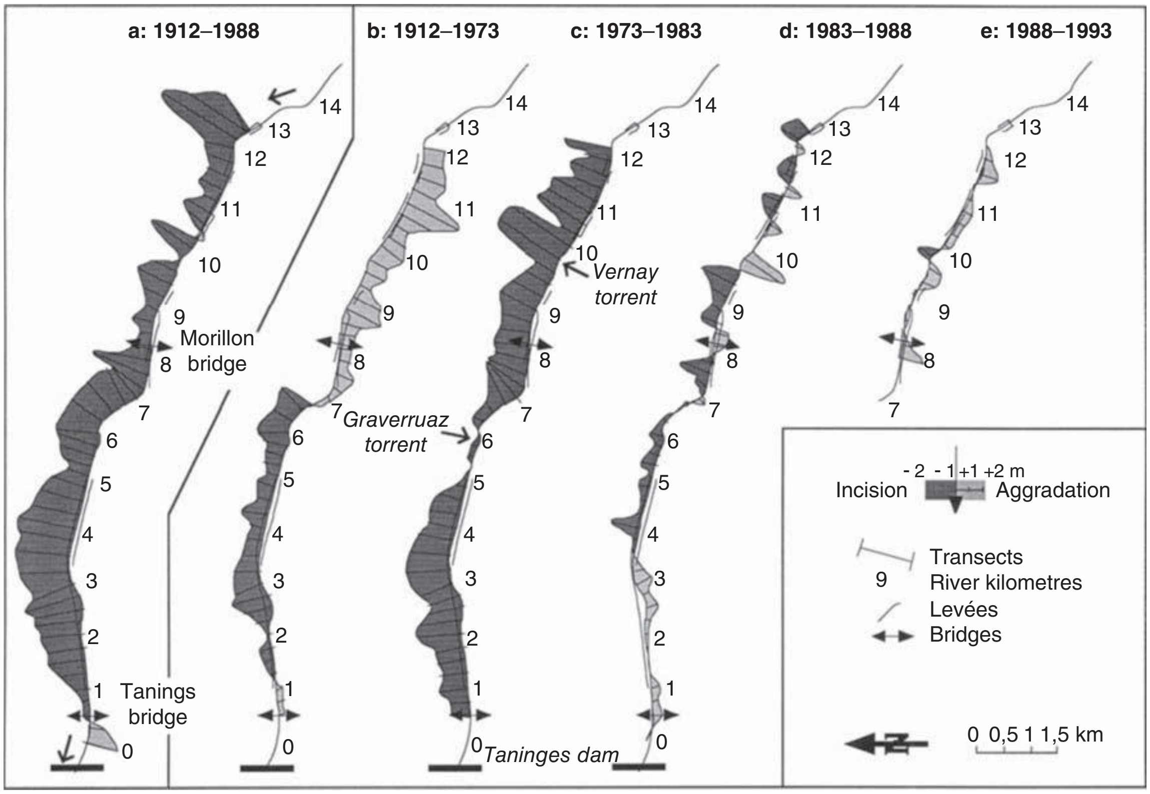


Figure 4.8 Long profile change on the Giffre River, French Alps. Source: Piegay and Peiry, 1997. Reproduced with permission of Elsevier.

river length). Positive and negative differences in altitude are shown by mapping the river line in plan and superimposing the deviations above and beneath the line (Landon and Piégay 1994; Piégay and Peiry 1997) (Fig. 4.8). Care needs to be taken to ensure that changes are not artefacts of differences in the spacing of survey points.

Bedload budgets can also be estimated from repeat long profile and/or cross profile information. The most accurate budgets are assessed when both types of data are available.

The superposition of sets of regularly spaced cross-sections allows accurate local assessment of changes in alluvial sediment storage and good-quality reach-scale assessments are also feasible if the cross-sections are closely spaced. For example, Vautier (1999) used five sets of cross-sections surveyed in 1949, 1965, 1972, 1984 and 1989 to establish channel degradation due to gravel mining and Surian and Cisotto (2007) used a set of 12 regularly surveyed cross-sections to document severe incision and channel narrowing over a 70 year period and to estimate gravel transport using a morphometric approach in the Brenta River, Italy. When using cross-section information for sediment budgeting, the cross-section resurveys need to relate to exactly the same location. If cross-sections are not resurveyed at exactly the same locations, the position of the channel floor needs to be interpolated before differences between the two surveys can be estimated, inevitably introducing significant errors.

Where only long profile information is available, a combination of changes in altitude derived from the long profiles and changes in channel width measured from maps or aerial photographs can be used to quantify bedload budgets at a reach scale and to study their spatial distribution, where approximate volumetric changes between two dates are calculated, reach by reach, using by the following equation:

$$V = \sum_{i=1}^{n} I_i \times L_i \times l_i$$

where V is the volume estimation (m^3), I_i the channel incision (m), L_i the length of the reach (m), l_i the channel width (m) and n the number of reaches.

In such analyses of cross and/or long profile data sets, changes in the elevation of the thalweg provide information on changes in the maximum channel depth and bedform position, but sediment budget estimation is best based on average bed elevations, for which frequent cross profiles are needed or water surface elevations.

Further information on constructing sediment budgets is given in Chapter 16.

Errors and uncertainty in the comparison of topographic records

Diachronic comparison of topographic records requires sets of comparable data. Unfortunately, several difficulties are

Table 4.3 Levelling networks in France from 1857 to present day.

Network name	Year of set-up	No. of polygons	Network length (km)	Altitude accuracy (cm km^{-1})	Altitude (centre of Paris) (m)	Difference in altitude (m)
Bardoulaouë	1857–1864	38	15000	±1.00	131.00	–
Lallemand	1884–1931	32	12715	±0.17	130.36	–0.64
IGN69	1963–1969	39	?	±0.13	130.70	+0.34

frequently met when researchers or engineers have to compare historical data. First, the reference system for altitude may have changed between survey dates. For example in France, three successive systems of levelling were set up from the middle of the 19th century (Table 4.3). Between 1857 and 1864, the building of the first railway lines and the extension of navigable canals led to the establishment of a first levelling network, which covered the whole country. The territory was covered by 38 polygons and zero altitude was the average level of the Mediterranean Sea at Marseille. On two later occasions, the network was changed through the replacement of geodetic landmarks and to increase the network accuracy (Landon 1999). Therefore, prior to any comparison of topographic records, it is essential to be sure that the altitude reference is identical for every set of data. In France, for example, maps are available for altitude conversion so that former values can be transposed to be compatible with the system used today. The conversion values are not constant in space, but increase from the South to the North, reaching a maximum +60 cm in northern France.

Second, a lack of data homogeneity is a serious obstacle to long and cross profile comparison. To avoid errors in geomorphological interpretation, it is preferable to compare topographical data of the same type, such as water surface levels with water surface levels, average bed levels with average bed levels, and so on. Long profiles constructed from average bed levels allow the most accurate comparisons. Long profiles of the water surface at low flows are strongly influenced by river discharge at the time of survey. The lack of discharge data for the time of survey is a frequent limitation to the use of this type of historical data, although in some cases water level–discharge relationships are available and can be used to correct the profile for this hydrological effect. Comparisons of thalweg profiles are relatively rare. Although these provide useful information on bedform change, their broader interpretation should be made carefully, because the migration of bedforms over time can lead to strong local variations in the thalweg elevation, which are independent of the general evolution of the river. Similar problems are relevant when comparisons of cross profiles from different survey dates are attempted. In addition, the precise relocation of cross profiles is essential because of the very large changes in channel cross-sectional form that occur within very short distances along a river.

Third, between two georeferenced points whose spatial location does not change over time (e.g. two bridges), the channel length may change with changes in river sinuosity. This is frequently the case on actively meandering rivers or on channels experiencing fluvial metamorphosis (e.g. from braiding to meandering). Under such circumstances, it becomes impossible to superimpose long profiles without first correcting the channel length. The best way to solve this problem is to calculate the ratio of channel sinuosity between the two dates and to then to adjust the horizontal distance scale along the long profile using this ratio. The ratio can be calculated reach-by-reach along a river valley, in order to ensure that the length correction is closely adapted to the local fluvial pattern. Under these circumstances, comparison of cross profile surveys also becomes problematic, not only because the channel position may have moved significantly, but also because the position of the cross profile with respect to the river's planform may have resulted in major cross profile changes that are not representative of wider adjustments in the channel. Overall resurveys of long or cross profiles of reaches that are showing changes in sinuosity and plan position should be undertaken with extreme care since apparent changes may simply be artefacts of survey locations relative to the channel planform.

Finally, errors in topographic survey data can be partitioned into the various sources and summed, as was discussed in relation to the cartographic record, to give an indication of positional accuracy. Errors related to the original survey methods, operator interpretation, geographical projection, location of the cross-section and, if the raw data are not available, the graphical or cartographic representation of the data can be accounted for at each point in time. Errors are propagated when surveys from different points in time are compared, so estimates of error for each point in time should be summed to give an indication of the minimum threshold for change detection.

4.5 The modern historical record: remote-sensing

The historical analysis of catchments and river channels is relying increasingly on remote sensing. As the time span covered by remotely sensed data increases, this resource is becoming, in effect, the modern historical record. In this section, we give a brief introduction to remote sensing and highlight its role in historical analysis using recent examples from the literature. Detailed discussion on the use of remotely sensed data in fluvial geomorphology is presented in Chapter 6.

Remote sensing approaches use instruments that are not in contact with the ground or water whose characteristics are being measured. Remote sensing methods may employ passive sensors that detect the electromagnetic radiation emanating from an object (e.g. photography) or active sensors that emit a signal and measure the properties of the signal after is has reflected off the object (e.g. radar), and the sensors may be mounted on satellites, aircraft or at points on the Earth's surface. Data types that are collected using remote sensing approaches include aerial photographs, multi- and hyperspectral data and radar- and laser-derived information (e.g. light detection and ranging, LiDAR).

The applicability of remotely sensed data to the study of temporal change in a river system is dependent on the spatial resolution of the data in relation to the size of the features and the magnitude of change being detected. For example, high-altitude (i.e. small-scale) aerial photography and most freely available satellite data have low spatial resolution, making them best suited to detect and quantify changes in large-scale features, such as land cover or the planform of large rivers (width $>\sim$100 m). Low-altitude aerial photography, airborne multispectral and LiDAR data sets have high spatial resolution, so can be employed to investigate changes in small-scale features, such as the planform characteristics or migration of narrow rivers or the characteristics of riparian vegetation.

Similarly, consideration must be given to the period of time during which the type of remotely sensed data has been collected and the frequency with which it is collected. Aerial photography is an excellent resource for historical analysis. Aerial photograph archives date back to the early to mid-20th century and in many areas photographs were taken routinely with at least a decadal frequency, which is sufficient to assess changes in most river systems. In general, satellite-based data sets span shorter periods, although Landsat has been monitoring the Earth's surface for over 40 years, but they benefit from more frequent data collection. Finally, high spatial resolution, hyperspectral and laser-derived data sets that are collected from airborne platforms or from terrestrial laser scanning (TLS) have only been collected over approximately the last decade.

Remotely sensed data can be employed in many types of historical analysis, including the assessment of temporal changes in catchment land cover/use, coarse sediment production and delivery to river systems, planform and channel migration, riparian vegetation and large wood dynamics, and the appraisal of geomorphic features in the channel and floodplain. Mirroring the earlier presentation of cartographic and topographic data, we present a few recent examples in which remotely sensed data has provided planimetric (two-dimensional) or volumetric (three-dimensional) information to investigations of historical change.

Most studies of planform change rely on aerial photographs, at least in part, for the interpretation of historical planform characteristics (e.g. Gurnell 1997b; Gaeuman *et al.* 2005; Bird *et al.* 2010; Nicoll and Hickin 2010; Michalkova *et al.* 2011; Moretto *et al.* 2013). Satellite multispectral data, particularly the long Landsat archive, have been used successfully to quantify temporal changes in planform for large rivers such as the Yellow, Ganges and Jamuna Rivers (Yao *et al.* 2011; Gupta *et al.* 2013; Mount *et al.* 2013). Aerial photographs and satellite data can also be used to examine the spatial coverage and temporal dynamics of channel geomorphic features. For example, Latrubesse *et al.* (2009) used Landsat imagery and aerial photography to assess changes in the frequency and size of bars and islands in the Araguaia River following catastrophic deforestation of the catchment.

Temporal changes in land cover/use can be assessed from aerial photographs and multispectral data. For example, Cadol *et al.* (2011) used aerial photographs to investigate the relationship between channel planform changes and the extent of woody riparian vegetation cover for 50 1 km reaches in Canyon de Chelly National Monument, United States, over a 70 year period. At a smaller scale, Meitzen (2009) used aerial imagery supported by field surveys to investigate changes in riparian vegetation structure and composition with lateral channel migration in the meandering Congaree River, United States. Henshaw *et al.* (2013) used multispectral data from Landsat to quantify changes in the position of channels and riparian vegetation for the braided Tagliamento, Italy, albeit with some limitations.

Volumetric changes in river channels and floodplains can be investigated using aerial photography and altimetry data from airborne and satellite-based sensors. Techniques using photogrammetry and spectral characteristics from aerial photography and multi/hyperspectral data are covered in Chapter 6, so here we focus instead on laser-based elevation data.

The basic technique for quantifying volumetric changes is known as DEM (digital elevation model) differencing, in which DEMs from different points in time are subtracted to determine the volumetric change over that period. The volumes being estimated can relate to any surface; land, water, bare gravel bars or riparian vegetation. The resulting three-dimensional data set is called a DEM of Difference (DoD). Although any topographic data can be used to construct DEMs and calculate DoDs, even historical topographic maps and surveys (James *et al.* 2012), the detailed spatial resolution and high vertical accuracy of LiDAR-derived DEMS make them ideally suited to examine changes in the topography of channel and floodplain surfaces over time. Much work has been conducted in recent years using LiDAR and TLS data to quantify volumetric change in river channels (e.g. Bowen and Waltermire 2002; Milan *et al.* 2007; De Rose and Basher 2011; O'Neal and Pizzuto 2011; Brasington *et al.* 2012), and also to estimate error in the DEMs and uncertainty in the change estimation (e.g. Wheaton *et al.* 2010; Milan *et al.* 2011). Recent applications by Bertoldi *et al.* (2011b, 2013) demonstrate the range of uses for LiDAR data. LiDAR data were used to investigate changes in bed morphology of the gravel-bed Tagliamento River and the extent of riparian forest. By combining these data with high oblique photography, field surveys and flow stage records, they were able to investigate the influence of riparian vegetation on the morphology of the braid plain

(Bertoldi *et al.* 2011b) and wood recruitment and deposition dynamics (Bertoldi *et al.* 2013). High-resolution DoDs can also be used to quantify changes in sediment production and delivery to the river associated with mass movements (DeLong *et al.* 2012 and references therein). In conclusion, LiDAR and TLS altimetry data provide unparalleled opportunities to investigate a range of questions related to temporal change in fluvial geomorphological forms and processes.

On a final note, recent technological advances in photogrammetry mean that elevation data can now be derived from any set of spatially overlapping photographs. Structure-from-motion (SfM) photogrammetry is an image-based method that can extract relative elevation data from digital images that bypasses some of the limitations of classical photogrammetry. Tests have shown that SfM can create high-density point clouds of topographic data that are comparable to LiDAR using only a digital camera and freely available software and for aerial photographs a tethered blimp or kite (Westoby *et al.* 2012; Fonstad *et al.* 2013). Significant issues remain to be resolved, particularly related to the estimation of error and uncertainty, but SfM is set to revolutionize and democratize the collection of high-resolution topographical data in fluvial geomorphology.

Accuracy and uncertainty

Considerable attention has been focused in recent years on approaches to assessing positional and attribute accuracy for remotely sensed data sets and to detect change over time. A thorough review is outside the scope of this chapter. Readers are referred to the earlier cartographic and topographic sections of this chapter for aspects that are relevant to remotely sensed data (e.g. spatial accuracy and digitization errors for planimetric measurements) and to recent texts (e.g. Congalton and Green 2009; Wheaton *et al.* 2010; Milan *et al.* 2011; Carbonneau and Piégay 2012).

4.6 Conclusion

Analysis of historical sources is often the only means to assess temporal variability in a river system. These sources allow us to peel back the layers of time to formulate an understanding of the past condition of a river and the pressures that have been and are currently influencing its form and processes and in turn to give us an indication of the possible future trajectories of the system. Thus, historical approaches are an important tool for the holistic management of river systems and an essential component in the decision-making and planning stages of river restoration measures.

The challenge of historical approaches is to assess and integrate information from a range of different sources that vary markedly in their reliability, accuracy and uncertainty. Through a systematic review of the historical data that are available for a site (Table 4.4), it is possible to evaluate the reliability of each source and to estimate its accuracy or uncertainty. Through the careful inclusion of historical data into a GIS, it becomes possible to compare directly different types of data that were collected at different scales and in relation to different map projections and to assess change over time inclusive of uncertainty and error. Also, through the use of catchment chronologies, it is possible to integrate data sources over time to explore qualitative and quantitative changes in catchment and river characteristics and to identify causal linkages. Only if the sources are fully understood and uncertainty is fully characterized can scientifically rigorous conclusions be drawn, otherwise it becomes impossible to differentiate genuine spatial or temporal patterns from those that are artefacts of the observer, recorder or cartographer.

Table 4.4 Steps in compiling historical data. Adapted from Hooke, 1997.

1	*Establish research sources and dates available*, and if possible use all available material including complementary archaeological and remote sensing sources
2	*Check background and general reliability* of sources
3	*Investigate document quality/accuracy/applicability*: source original documents or verify compilations based upon secondary sources; note purpose of records; undertake content analysis if appropriate
4	*Investigate topographic survey quality/accuracy/applicability*: check accuracy of individual surveys, including planimetric accuracy and content accuracy, and identify points of common detail to enable comparison of different surveys
5	*Create time sequence of catchment, river or reach conditions* using both qualitative and quantitative methods as appropriate
6	*Field check changes* indicated

Finally, information extracted from historical sources should not be seen in isolation. A combination of contemporary data, particularly field survey data relating to, for example, bed and bank sediment calibre and structure and riparian vegetation structure and age, can add new dimensions to the historical analysis, helping to extend and validate interpretations based on historical sources (e.g. Bertoldi *et al.* 2011a, 2013; Rollet *et al.* 2013).

Acknowledgements

The authors would like to acknowledge and thank J.L. Peiry and G.E. Petts for their contributions to the first edition, some of which has been transfered to this updated chapter. Research on the use of historical sources conducted within the REFORM collaborative project funded by the European Union Seventh Framework Programme under grant agreement 282656 has contributed greatly to the revision of this chapter.

References

Bender, O., Boehmer, H.J., Jens, D. and Schumacher, K.P. 2005. Using GIS to analyse long-term cultural landscape change in Southern Germany. *Landscape and Urban Planning* 70(1–2): 111–125.

Berglund, B.E., ed. 1986. *Handbook of Holocene Palaeoecology and Palaeohydrology*, Chichester: John Wiley & Sons.

Bertoldi, W., Drake, N.A. and Gurnell, A.M. 2011a. Interactions between river flows and colonizing vegetation on a braided river: exploring spatial and temporal dynamics in riparian vegetation cover using satellite data. *Earth Surface Processes and Landforms* 36: 1474–1486.

Bertoldi, W., Gurnell, A.M. and Drake, N.A. 2011b. The topographic signature of vegetation development along a braided river: results of a combined analysis of airborne lidar, colour air photographs and ground measurements. *Water Resources Research* 47: W06525.

Bertoldi, W., Gurnell, A.M. and Welber, M. 2013. Wood recruitment and retention: the fate of eroded trees on a braided river explored using a combination of field and remotely-sensed data sources. *Geomorphology* 180: 146–155.

Beschta, R.L. and Ripple, W.J. 2006. River channel dynamics following extirpation of wolves in northwestern Yellowstone National Park, USA. *Earth Surface Processes and Landforms* 31(12): 1525–1539.

Bird, S., Hogan, D. and Schwab, J. 2010. Photogrammetric monitoring of small streams under a riparian forest canopy. *Earth Surface Processes and Landforms* 35(8): 952–970.

Blanic, R. and Verdet, G. 1975. Quelques travaux de correction sur le cours de l'Isère. *La Houille Blanche* 2–3: 191–197.

Boix-Fayos, C., Barbera, G.G., Lopez-Bermudez, F. and Castillo, V.M. 2007. Effects of check dams, reforestation and land-use changes on river channel morphology: case study of the Rogativa catchment (Murcia, Spain). *Geomorphology* 91(1–2): 103–123.

Bowen, Z.H. and Waltermire, R.G. 2002. Evaluation of light detection and ranging (LIDAR) for measuring river corridor topography. *Journal of the American Water Resources Association* 38(1): 33–41.

Bradley, R.S. and Jones, P.D. 1992. *Climate Since A.D. 1500*, London: Routledge.

Brasington, J., Vericat, D. and Rychkov, I. 2012. Modeling river bed morphology, roughness, and surface sedimentology using high resolution terrestrial laser scanning. *Water Resources Research*, 48: W11519.

Bravard, J.P. 1987. Le Rhône du Léman à Lyon. In: Michel, C., ed., *Revue de Géographie Alpine*, Grenoble: Institut de Géographie Alpine, pp. 89–91.

Bravard, J.P. 1994. La charge de fond du Haut-Rhône français, mise en perspective historique. *Dossier de la Revue de Géographie Alpine* 12: 163–169.

Brooks, A.P. and Brierley, G.J. 1997. Geomorphic responses of lower Bega River to catchment disturbance, 1851–1926. *Geomorphology* 18(3–4): 291–304.

Brown, A.G. 2009. Colluvial and alluvial response to land use change in Midland England: an integrated geoarchaeological approach. *Geomorphology* 108(1–2): 92–106.

Burt, T.P. and Gardiner, A.T. 1982. The permanence of stream networks in Britain – some further comments. *Earth Surface Processes and Landforms* 7(4): 327–332.

Burt, T.P. and Oldman, J.C. 1986. The permanence of stream networks in Britain – further comments. *Earth Surface Processes and Landforms* 11(1): 111–113.

Butlin, R.A. 1990. Drainage and landuse in the Fenlands and Fen-edge of northeast Cambridgeshire in the seventeenth and eighteenth centuries. In: Cosgrove, D.E. and Petts, G.E., eds., *Water, Engineering and Landscape: Water Control and Landscape Transformation in the Modern Period*, London: Belhaven Press, pp. 54–76.

Cadol, D., Rathburn, S.L. and Cooper, D.J. 2011. Aerial photographic analysis of channel narrowing and vegetation expansion in Canyon de Chelly National Monument, Arizona, USA 1935–2004. *River Research and Applications* 27(7): 841–856.

Calver, E.K. 1853. *Conservation and Improvement of Tidal Rivers*, London: John Weale.

Camden, J. 1586. *Britannia*: citation is from the Edmund Gibson 1695 ed., from the facsimile edited by S. Piggott, Newton Abbott (1971), cited in Butlin, R.A. (1990), Drainage and landuse in the Fenlands and Fen-edge of northeast Cambridgeshire in the seventeenth and eighteenth centuries. In: Cosgrove, D.E. and Petts, G.E., eds., *Water, Engineering and Landscape: Water Control and Landscape Transformation in the Modern Period*, London: Belhaven Press, pp. 54–76.

Carbonneau, P. and Piégay, H. 2012. *Fluvial Remote Sensing for Science and Management*, Oxford: Wiley-Blackwell.

Carr, A.P. 1962. Cartographie record and historical accuracy. *Geography* 47(2): 135–144.

Cheung, C.K. and Shi, W.Z. 2004. Estimation of the positional uncertainty in line simplification in GIS. *Cartographic Journal* 41(1): 37–45.

Chrisman, N.R. 1991. The error component in spatial data. In: Goodchild, M.F. and Gopal, S., eds., *Geographic Information Systems. Principles and Applications*, London: Longman, pp. 165–174.

Collins, B. and Dunne, T. 1989. Gravel transport, gravel harvesting, and channel-bed degradation in rivers draining the southern olympic mountains, Washington, U.S.A. *Environmental Geology and Water Sciences* 13(3): 213–224.

Comiti, F., Da Canal, M., Surian, N., Mao, L., Picco, L. and Lenzi, M.A. 2011. Channel adjustments and vegetation cover dynamics in a large gravel bed river over the last 200 years. *Geomorphology* 125(1): 147–159.

Congalton, R.G. and Green, K. 2009. *Assessing the Accuracy of Remotely Sensed Data : Principles and Practices*, Boca Raton, FL: CRC Press, 183pp.

Cooke, R.U. and Reeves, R.W. 1976. *Arroyos and Environmental Change in the American South-West*, Oxford: Clarendon Press,.

Davis, M.B. 1976. Erosion rates and land-use history in southern Michigan. *Environmental Conservation* 3(2): 139–148.

De Jager, N.R., Rohweder, J.J. and Nelson, J.C. 2013. Past and predicted future changes in the land cover of the Upper Mississippi River floodplain, USA. *River Research and Applications* 29(5): 608–618.

DeLong, S.B., Prentice, C.S., Hilley, G.E. and Ebert, Y. 2012. Multitemporal ALSM change detection, sediment delivery, and process mapping at an active earthflow. *Earth Surface Processes and Landforms* 37(3): 262–272.

Del Rio, L. and Javier Gracia, F. 2013. Error determination in the photogrammetric assessment of shoreline changes. *Natural Hazards* 65(3): 2385–2397.

De Rose, R.C. and Basher, L.R. 2011. Measurement of river bank and cliff erosion from sequential LIDAR and historical aerial photography. *Geomorphology* 126(1–2): 132–147.

Downs, P.W., Dusterhoff, S.R. and Sears, W.A. 2013. Reach-scale channel sensitivity to multiple human activities and natural events: Lower Santa Clara River, California, USA. *Geomorphology* 189: 121–134.

Dugdale, W. 1772. *History of Imbanking and Draining of Divers Fens and Marshes Both in Foreign Parts and in this Kingdom* London: Geast,.

Ellett, C. 1853. *The Mississippi and Ohio Rivers*, Philadelphia, PA: Lippincott, Grambo & Co.

Emmett, W.W. and Hadley, R.F. 1968. *The Vigil Network – Preservation and Access of Data, US Geological Survey Circular 460-C*, Washingtoin, DC: US Geological Survey, 21pp.

Erskine, W.D. 1992. Channel response to large-scale river training works – Hunter River, Australia. *Regulated Rivers – Research & Management* 7(3): 261–278.

Erskine, W.D. 1999. Oscillatory response versus progressive degradation of incised channels in Southeastern Australia. In: Darby, S.E. and Simon, A., eds., *Incised River Channels : Processes, Forms, Engineering and Management*, Chichester: John Wiley & Sons.

Everett, B.L. 1968. Use of cottonwood in the investigation of the recent history of a flood plain. *American Journal of Science* 206: 417–439.

Fisher, P. 1991. Spatial data sources and data problems. In: Maguire, D.G., Goodchild, M.F. and Rhind, D.W., eds., *Geographical Information Systems. Principles and Applications*, vol. 1, Harlow: Longman, pp. 175–189.

Flowerdew, R. 1991. Spatial data integration. In: Maguire, D.G., Goodchild, M.F. and Rhind, D.W., eds., *Geographical Information Systems. Principles and Applications*, vol. 1, Harlow: Longman, pp. 375–387.

Fonstad, M.A., Dietrich, J.T., Courville, B.C., Jensen, J.L. and Carbonneau, P.E. 2013. Topographic structure from motion: a new development in photogrammetric measurement. *Earth Surface Processes and Landforms* 38(4): 421–430.

Foster, I.D.L., Dearing, J.A., Simpson, A., Carter, A.D. and Appleby, P.G. 1985. Lake catchment based studies of erosion and denudation in the merevale catchment, Warwickshire, U.K. *Earth Surface Processes and Landforms* 10(1): 45–68.

Foster, I.D.L., Collins, A.L., Naden, P.S., Sear, D.A., Jones, J.I. and Zhang, Y. 2011. The potential for paleolimnology to determine historic sediment delivery to rivers. *Journal of Paleolimnology* 45(2): 287–306.

Gaeuman, D., Schmidt, J.C. and Wilcock, P.R. 2005. Complex channel responses to changes in stream flow and sediment supply on the lower Duchesne River, Utah. *Geomorphology* 64(3–4): 185–206.

Galatowitsch, S.M. 1990. Using the original land survey notes to reconstruct presettlement landscapes in the American West. *The Great Basin Naturalist* 50(2): 181–191.

Gardiner, V. 1975. Drainage basin morphometry. *British Geomorphological Research Group, Technical Bulletin*, 14.

Gemaehling, C. and Chabert, J. 1962. Transport solide et modification du lit d'une rivière à forte pente: la Drôme. In: *Colloque de Bari*, Publication 59, Gentbrugge: IAHS – Association Internationale d'Hydrologie Scientifique, pp. 259–272.

Goldberg, P. and Macphail, R. 2006. *Practical and Theoretical Geoarchaeology*, Malden, MA: Blackwell.

Greco, S.E., Fremier, A.K., Larsen, E.W. and Plant, R.E. 2007. A tool for tracking floodplain age land surface patterns on a large meandering river with applications for ecological planning and restoration design. *Landscape and Urban Planning* 81(4): 354–373.

Gregory, K.J. 1983. *Background to Palaeohydrology : a Perspective*, Chichester: John Wiley & Sons.

Gregory, K.J., Lewin, J. and Thornes, J.B. 1987. *Palaeohydrology in Practice: a River Basin Analysis*, Chichester: John Wiley & Sons..

Gupta, N., Atkinson, P.M. and Carling, P.A. 2013. Decadal length changes in the fluvial planform of the River Ganga: bringing a mega-river to life with Landsat archives. *Remote Sensing Letters* 4(1): 1–9.

Gurnell, A.M. 1996. Hydrological and geomorphological change on the River Dee meanders. In: Maxwell, W.H.C., Preul, H.C. and Stout, G.E., eds., *Rivertech '96 – 1st International Conference on New/Emerging Concepts for Rivers, Proceedings*, vol. 2, Urbana, IL: International Water Resources Association, pp. 484–491.

Gurnell, A.M. 1997a. Adjustments in river channel geometry associated with hydraulic discontinuities across the fluvial–tidal transition of a regulated river. *Earth Surface Processes and Landforms* 22(10): 967–985.

Gurnell, A.M. 1997b. Channel change on the River Dee meanders 1946–1992, from the analysis of air photographs. *Regulated Rivers – Research & Management* 13(1): 13–26.

Gurnell, A.M., Downward, S.R. and Jones, R. 1994. Channel planform change on the River Dee meanders, 1876–1992. *Regulated Rivers – Research & Management* 9(4): 187–204.

Harley, J.B. 1975. *Ordnance Survey Maps: a Descriptive Manual*. Southampton: Ordnance Survey.

Harley, J.B. 1982. Historical geography and its evidence: reflections on modelling sources. In: Baker, A.R.H. and Billinge, M., eds., *Period and Place: Research Methods in Historical Geography*, Cambridge: Cambridge University Press, pp. 261–273.

Harmar, O.P. and Clifford, N.J. 2006. Planform dynamics of the Lower Mississippi River. *Earth Surface Processes and Landforms* 31(7): 825–843.

Henshaw, A.J., Gurnell, A.M., Bertoldi, W. and Drake, N.A. 2013. An assessment of the degree to which Landsat TM data can support the assessment of fluvial dynamics, as revealed by changes in vegetation extent and channel position, along a large river. *Geomorphology* 202: 74–85.

Hohensinner, S., Habersack, H., Jungwirth, M. and Zauner, G. 2004. Reconstruction of the characteristics of a natural alluvial river–floodplain system and hydromorphological changes following human modifications: the Danube River (1812–1991). *River Research and Applications* 20(1): 25–41.

Hohensinner, S., Jungwirth, M., Muhar, S. and Schmutz, S. 2011. Spatio-temporal habitat dynamics in a changing Danube River landscape 1812–2006. *River Research and Applications* 27(8): 939–955.

Hohensinner, S., Lager, B., Sonnlechner, C., Haidvogl, G., Gierlinger, S., Schmid, M., Krausmann, F. and Winiwarter, V. 2013a. Changes in water and land: the reconstructed Viennese riverscape from 1500 to the present. *Water History* 5(2): 145–172.

Hohensinner, S., Sonnlechner, C., Schmid, M. and Winiwarter, V. 2013b. Two steps back, one step forward: reconstructing the dynamic Danube riverscape under human influence in Vienna. *Water History* 5(2): 121–143.

Hooke, J.M. 1997. Styles of channel change. In: Hey, R.D., Thorne, C.R. and Newson, M.D., eds., *Applied Fluvial Geomorphology for River Engineering and Management*, Chichester: John Wiley & Sons, pp. 237–268.

Hooke, J.M. 2007. Spatial variability, mechanisms and propagation of change in an active meandering river. *Geomorphology* 84(3–4): 277–296.

Hooke, J.M. and Kain, R.J.P. 1982. *Historical Change in the Physical Environment: a Guide to Sources and Techniques*, London: Butterworth.

Hooke, J.M. and Redmond, C.E. 1989. Use of cartographic sources for analysing river channel change with examples from Britain. In: Petts, G.E., Möller, H. and Roux, A.L., eds., *Historical Change of Large Alluvial Rivers: Western Europe*, Chichester: John Wiley & Sons, pp. 79–93.

Hu, B. 2010. Application of geographic information systems (GIS) in the history of cartography. *World Academy of Science, Engineering and Technology*, 42: 185–189.

Hudson, P.F. and Kesel, R.H. 2000. Channel migration and meander-bend curvature in the lower Mississippi River prior to major human modification. *Geology* 28(6): 531–534.

James, L.A., Hodgson, M.E., Ghoshal, S. and Latiolais, M.M. 2012. Geomorphic change detection using historic maps and DEM differencing: the temporal dimension of geospatial analysis. *Geomorphology* 137(1): 181–198.

Jessop, W. 1782. *Report of William Jessop, Engineer, on a Survey of the River Trent, in the Months of August and September 1782, Relative to a Scheme for Improving Its Navigation*, Nottingham: Burbage and Son.

Kochel, R.C. and Baker, V.R. 1988. Palaeoflood analysis using slackwater deposits. In: Baker, V.R., Kochel, R.C. and Patton, P.C., eds., *Flood Geomorphology*, New York: John Wiley & Sons, pp. 357–376.

Kondolf, G.M. and Larson, M. 1995. Historical channel analysis and its application to riparian and aquatic habitat restoration. *Aquatic Conservation – Marine and Freshwater Ecosystems* 5(2): 109–126.

Kondolf, G.M. and Piégay, H. 2007. Changes in the riparian zone of the lower Eygues River, France, since 1830. *Landscape Ecology* 22: 367–384.

Kondolf, G.M. and Swanson, M.L. 1993. Channel adjustments to reservoir construction and gravel extraction along Stony Creek, California. *Environmental Geology* 21(4): 256–269.

Kummu, M., Lu, X.X., Rasphone, A., Sarkkula, J. and Koponen, J. 2008. Riverbank changes along the Mekong River: remote sensing detection in the Vientiane–Nong Khai area. *Quaternary International* 186: 100–112.

Landon, N. 1999. *L'Évolution Contemporaine du Profil en Long des Afffluents du Rhône Moyen. Constat Régional et Analyse d'un Hydrosystème Complexe, la Drôme*, unpublished Thesis in Geography, Université Paris 4 – Sorbonne.

Landon, N. and Piégay, H. 1994. L'incision d'affluents méditerranéens du Rhône: la Drôme et l'Ardèche. *Revue de Géographie de Lyon* 69(1): 63–72.

Large, A.R.G. and Petts, G.E. 1996. Historical channel–floodplain dynamics along the River Trent – implications for river rehabilitation. *Applied Geography* 16(3): 191–209.

Latrubesse, E.M., Amsler, M.L., de Morais, R.P. and Aquino, S. 2009. The geomorphologic response of a large pristine alluvial river to tremendous deforestation in the South American tropics: the case of the Araguaia River. *Geomorphology* 113(3–4): 239–252.

Library of Congress 2006. *Technical Guidelines for Digital Conversion of Text and Graphic Materials*, Washington, DC: US Library of Congress. http://memory.loc.gov/ammem/about/techStandards.pdf (accessed 5 November 2013).

Manies, K.L. and Mladenoff, D.J. 2000. Testing methods to produce landscape-scale presettlement vegetation maps from the US public land survey records. *Landscape Ecology* 15(8): 741–754.

Manzano-Agugliaro, F., San-Antonio-Gómez, C., López, S., Montoya, F.G. and Gil, C. 2013. Pareto-based evolutionary algorithms for the calculation of transformation parameters and accuracy assessment of historical maps. *Computers & Geosciences* 57: 124–132.

Martin-Vide, J.P., Ferrer-Boix, C. and Ollero, A. 2010. Incision due to gravel mining: modeling a case study from the Gallego River, Spain. *Geomorphology* 117(3–4): 261–271.

Maser, C. and Sedell, J.R. 1994. *From the Forest to the Sea: the Ecology of Wood in Streams, Rivers, Estuaries, and Oceans*, Delray Beach, FL: St. Lucie Press, 200pp.

Meitzen, K.M. 2009. Lateral channel migration effects on riparian forest structure and composition, Congaree River, South Carolina, USA. *Wetlands* 29(2): 465–475.

Michalkova, M., Piégay, H., Kondolf, G.M. and Greco, S.E. 2011. Lateral erosion of the Sacramento River, California (1942–1999), and responses of channel and floodplain lake to human influences. *Earth Surface Processes and Landforms* 36(2): 257–272.

Milan, D.J., Heritage, G.L. and Hetherington, D. 2007. Application of a 3D laser scanner in the assessment of erosion and deposition volumes and channel change in a proglacial river. *Earth Surface Processes and Landforms* 32(11): 1657–1674.

Milan, D.J., Heritage, G.L., Large, A.R.G. and Fuller, I.C. 2011. Filtering spatial error from DEMs: Implications for morphological change estimation. *Geomorphology* 125(1): 160–171.

Milne, G. and Hobley, B., eds. 1981. *Waterfront Archaeology in Britain and Northern Europe, CBA Research Report 41*, York: Council for British Archaeology.

Moretto, J., Rigon, E., Mao, L., Picco, L., Delai, F. and Lenzi, M.A. 2013. Channel adjustments and island dynamics in the Brenta River (Italy) over the last 30 years. *River Research and Applications* 30(6): 719–732.

Mount, N. and Louis, J. 2005. Estimation and propagation of error in measurements of river channel movement from aerial imagery. *Earth Surface Processes and Landforms* 30(5): 635–643.

Mount, N.J., Louis, J., Teeuw, R.M., Zukowskyj, P.M. and Stott, T. 2003. Estimation of error in bankfull width comparisons from temporally sequenced raw and corrected aerial photographs. *Geomorphology* 56(1–2): 65–77.

Mount, N.J., Tate, N.J., Sarker, M.H. and Thorne, C.R. 2013. Evolutionary, multi-scale analysis of river bank line retreat using continuous wavelet transforms: Jamuna River, Bangladesh. *Geomorphology* 183: 82–95.

Nicoll, T.J. and Hickin, E.J. 2010. Planform geometry and channel migration of confined meandering rivers on the Canadian prairies. *Geomorphology*, 116(1–2): 37–47.

O'Neal, M.A. and Pizzuto, J.E. 2011. The rates and spatial patterns of annual riverbank erosion revealed through terrestrial laser-scanner surveys of the South River, Virginia. *Earth Surface Processes and Landforms* 36(5): 695–701.

Osterkamp, W.R., Emmett, W.W. and Leopold, L.B. 1991. The Vigil Network: a means of observing landscape change in drainage basins. *Hydrological Sciences Journal* 36(4): 331–344.

Ovenden, J.C. and Gregory, K.J. 1980. The permanence of stream networks in Britain. *Earth Surface Processes* 5(1): 47–60.

Patrick, D.M., Smith, L.M. and Whitten, C.B. 1982. Methods for studying accelerated fluvial change. In: Hey, R.D., Bathurst, J.C. and Thorne, C.R., eds., *Gravel-bed Rivers: Fluvial Processes, Engineering and Management*, Chichester: John Wiley & Sons, pp. 783–815.

Peiry, J.L. 1987. Channel degradation in the middle Arve river, France. *Regulated Rivers – Research & Management* 1(2): 183–188.

Petts, G.E. 1979. Complex response of river channel morphology subsequent to reservoir construction. *Progress in Physical Geography* 3(3): 329–362.

Petts, G.E. 1995. Changing river channels: the geographical tradition. In: Gurnell, A.M. and Petts, G.E., eds., *Changing River Channels*, Chichester: John Wiley & Sons, pp. 1–23.

Petts, G.E. and Pratts, J.D. 1983. Channel changes following reservoir construction on a Lowland English River. *Catena* 10(1–2): 77–85.

Petts, G.E., Möller, H. and Roux, A.L., eds. 1989. *Historical Change of Large Alluvial Rivers: Western Europe*, Chichester: John Wiley & Sons, 355pp.

Pfister, C. 1992. Monthly temperature and precipitation in central Europe from 1525–1979: quantifying documentary evidence on weather and its effects. In: Bradley, R.S. and Jones, P.D., eds., *Climate Since A.D. 1500*, London: Routledge, pp. 118–142.

Piégay, H. and Peiry, J.L. 1997. Long profile evolution of a mountain stream in relation to gravel load management: example of the middle Giffre River (French Alps). *Environmental Management* 21(6): 909–919.

Piégay, H., Darby, S.E., Mosselman, E. and Surian, N. 2005. A review of techniques available for delimiting the erodible river corridor: a sustainable approach to managing bank erosion. *River Research and Applications* 21(7): 773–789.

Piégay, H., Alber, A., Slater, L. and Bourdin, L. 2009. Census and typology of braided rivers in the French Alps. *Aquatic Sciences* 71(3): 371–388.

Rinaldi, M. 2003. Recent channel adjustments in alluvial rivers of Tuscany, central Italy. *Earth Surface Processes and Landforms* 28(6): 587–608.

Rinaldi, M. and Simon, A. 1998. Bed-level adjustments in the Arno River, central Italy. *Geomorphology* 22(1): 57–71.

Rollet, A.J., Piégay, H., Dufour, S., Bornette, G. and Persat, H. 2013. Assessment of consequences of sediment deficit on a gravel river bed downstream of dams in restoration perspectives: application of a multicriteria, hierarchical and spatially explicit diagnosis. *River Research and Applications* 30(8): 939–953.

Sear, D.A., Darby, S.E., Thorne, C.R. and Brookes, A.B. 1994. Geomorphological approach to stream stabilization and restoration – case study of the Mimmshall Brook, Hertfordshire, UK. *Regulated Rivers – Research & Management* 9(4): 205–223.

Smelser, M.G. and Schmidt, V.C. 1998. *An Assessment Methodology for Determining Historical Changes in Mountain Streams, USDA Forest Service General Technical Report RMRS-GTR-6*, Fort Collins, CO: Rocky Mountain Research Station.

Snell, F.C. 1938. *The Intermittent, or Nailbourne, Streams of East Kent*, Canterbury: Hunt, Snell and Co.

Starkel, L., Gregory, K.J. and Thornes, J.B. 1991. *Temperate Palaeohydrology: Fluvial Processes in the Temperate Zone During the Last 15,000 Years*, New York: John Wiley & Sons.

Surian, N. 1999. Channel changes due to river regulation: the case of the Piave River, Italy. *Earth Surface Processes and Landforms* 24(12): 1135–1151.

Surian, N. and Cisotto, A. 2007. Channel adjustments, bedload transport and sediment sources in a gravel-bed river, Brenta River, Italy. *Earth Surface Processes and Landforms* 32(11): 1641–1656.

Surian, N., Ziliani, L., Comiti, F., Lenzi, M.A. and Mao, L. 2009. Channel adjustments and alteration of sediment fluxes in gravel-bed rivers of north-eastern Italy: potentials and limitations for channel recovery. *River Research and Applications* 25(5): 551–567.

Trimble, S.W. 1998. Dating fluvial processes from historical data and artifacts. *Catena* 31(4): 283–304.

Trimble, S.W. 2008. The use of historical data and artifacts in geomorphology. *Progress in Physical Geography* 32(1): 3–29.

Trimble, S.W. 2012. Historical sources and watershed evolution. *Philosophical Transactions of the Royal Society A – Mathematical Physical and Engineering Sciences*, 370(1966): 2075–2092.

Trimble, S.W. and Carey, W.P. 1984. A comparison of the Brune and Churchill methods for computing sediment yields applied to a reservoir system. In: Subitsky, S, ed., *Selected Papers in the Hydrologic Sciences, 1988–1992, US Geological Survey Water Supply Paper 2340*, Denver, CO: US Geological Survey, pp. 195–202.

Tucci, M. and Giordano, A. 2011. Positional accuracy, positional uncertainty, and feature change detection in historical maps: results of an experiment. *Computers, Environment and Urban Systems* 35(6): 452–463.

Uribelarrea, D., Perez-Gonzalez, A. and Benito, G. 2003. Channel changes in the Jararna and Tagus rivers (central Spain) over the past 500 years. *Quaternary Science Reviews* 22(20): 2209–2221.

Vautier, F. 1999. *Dynamique Fluviale et Végétalisation des Cours d'Eau Alpins Endigués: l'Exemple de l'Isère dans la Vallée du Grésivaudan*, Thesis in Geography, Université Joseph Fourier, Grenoble.

Walter, R.C. and Merritts, D.J. 2008. Natural streams and the legacy of water-powered mills. *Science* 319(5861): 299–304.

Westoby, M.J., Brasington, J., Glasser, N.F., Hambrey, M.J. and Reynolds, J.M. 2012. 'Structure-from-notion' photogrammetry: a low-cost, effective tool for geoscience applications. *Geomorphology* 179(0): 300–314.

Wheaton, J.M., Brasington, J., Darby, S.E. and Sear, D.A. 2010. Accounting for uncertainty in DEMs from repeat topographic surveys: improved sediment budgets. *Earth Surface Processes and Landforms* 35(2): 136–156.

Williams, G.P. and Wolman, M.G. 1984. *Downstream Effects of Dams on Alluvial Rivers, US Geological Survey Professional Paper 1286*, Washington, DC: US Government Printing Office.

Winterbottom, S.J. 2000. Medium and short-term channel planform changes on the Rivers Tay and Tummel, Scotland. *Geomorphology* 34(3–4): 195–208.

Wishart, D., Warburton, J. and Bracken, L. 2008. Gravel extraction and planform change in a wandering gravel-bed river: the River Wear, Northern England. *Geomorphology* 94(1–2): 131–152.

Yao, Z., Ta, W., Jia, X. and Xiao, J. 2011. Bank erosion and accretion along the Ningxia–Inner Mongolia reaches of the Yellow River from 1958 to 2008. *Geomorphology* 127(1–2): 99–106.

Ziliani, L. and Surian, N. 2012. Evolutionary trajectory of channel morphology and controlling factors in a large gravel-bed river. *Geomorphology* 173–174: 104–117.

SECTION III

The Spatial Framework: Emphasizing Spatial Structure and Nested Character of Fluvial Forms

CHAPTER 5

System approaches in fluvial geomorphology

Hervé Piégay
Université de Lyon, UMR 5600 CNRS, Lyon, France

5.1 System, fluvial system, hydrosystem

The system, a widespread concept

The system concept, increasingly used in environmental sciences during the last four decades to link physical, chemical and biotic processes, has had an important influence on fluvial geomorphology (Hack 1960; Chorley and Kennedy 1971; Schumm 1977; Bennett and Chorley 1978; Huggett 2007). As links among geomorphology, its sister disciplines (ecology, hydrology and human geography) and river management have developed, they have strongly influenced fluvial geomorphology, notably with articulation of the 'hydrosystem' (Amoros *et al.* 1982, 1988; Amoros and Petts 1993) and in Australia, the concept of 'river styles' (Brierley and Fryirs 2005; Brierley *et al.* 2008). While equilibrium concepts in geomorphology were arguably inspired by classical mechanics and thermodynamics, notions of dynamic equilibrium can be seen as having tracked developments in open-state thermodynamics (Huggett 2007).

The concept of a system is a tool in the sense that it is used to organize research. Although providing important insights into processes, reductionist approaches typically cannot bring a general understanding of rivers and their evolution. In this context, the system concept provides a framework to develop an integrated picture of geomorphic processes and forms on larger time and spatial scales, which have appeal for river managers who seek to implement the concept of 'sustainable development' and to better integrate scientific insights into management.

The fluvial system

A system can be defined as a meaningful combination of elements that form a complex whole, with connections, interrelations and transfers of energy and matter among them. The term fluvial derives from the Latin word *fluvius*, a river, but when carried to its broadest interpretation, a fluvial system not only involves stream channels but also entire drainage networks and depositional zones of deltas and alluvial fans and also to the hillslope sources of runoff and sediments.

The fluvial system is a complex adaptive process–response system with two main physical components, the morphological system of channels, floodplains, hillslopes, deltas, etc., and the cascading system of the flow of water and sediment (Chorley and Kennedy 1971). The fluvial system changes progressively through geological time, as a result of normal erosional and depositional processes, and it responds to changes of climate, base level, tectonics and human impacts (Fig. 5.1). Hence there can be considerable variability of fluvial system morphology and dynamics through time from natural processes alone. In addition to this, since at least the beginning of the Neolithic, human activities have played a major role in fluvial system evolution, affecting vegetation cover, base level, and also water, sediment and organic matter inputs on time-scales that may be very short compared with those on which climate and tectonic changes are usually acting (Park 1981; Gregory 1987). This human influence is now so important that a new geological epoch has been proposed, the 'Anthropocene', from 1800 to present (Meybeck 2003). In addition to this temporal variability, there is a strong spatial variability resulting from different geological, climatic, topographic and societal environments. The prediction and postdiction of fluvial system behaviour is greatly complicated by this variability (Fig. 5.1).

At the channel scale, we conventionally summarize the fluvial system as a set of variables, some being the control/external/independent variables (e.g. Q_s, the sediment load, and Q_p, the peak discharge) and the others are the adjustable/internal/dependent variables (e.g. channel pattern, meander wavelength, channel slope, width and depth). The river is then seen to be in a dynamic equilibrium when the adjustable variables vary slightly around an average through time. When the control factors change, the fluvial system undergoes a correlative change, the dependent variables adjusting to a new equilibrium. This stage is called a fluvial metamorphosis and is often illustrated in the scientific literature by a planform change.

Some systems adjust rapidly to changes within the basin, whereas others are more resistant. The thresholds for change (i.e. an erosional or depositional adjustment) vary from one system to another. Within the French Alps, intensive instream gravel mining in the 1970s and 1980s induced channel degradation for kilometres upstream and downstream of the mining sites. In reaches where shallow (~4 m) gravel layers are underlain by fine lacustrine deposits of the post-Würmian glaciation,

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

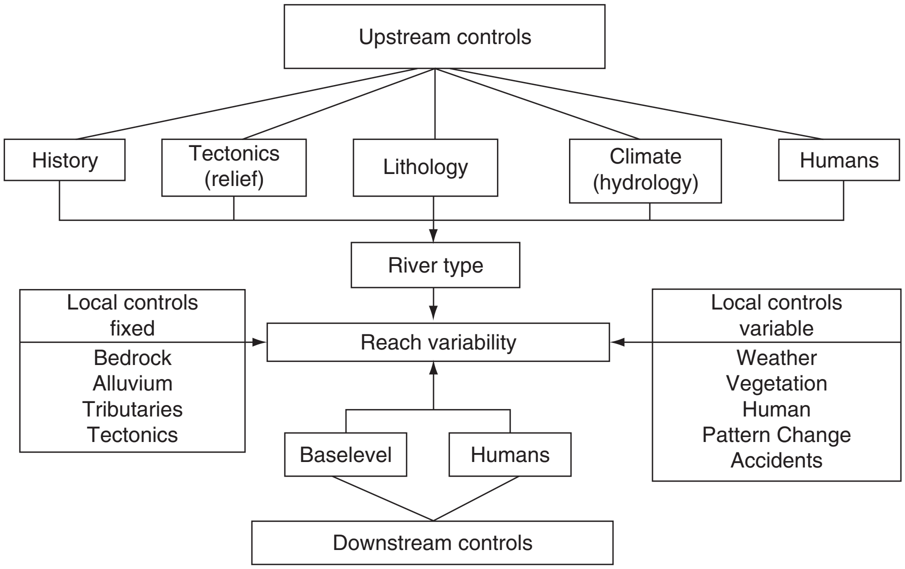


Figure 5.1 The fluvial system: a conceptual model of the geomorphological functioning of a river basin focusing on the channel reach variability associated with upstream, local and downstream controls.

once the gravel was gone and lacustrine silts exposed, incision was rapid, reaching 10 m over many reaches (Bravard *et al.* 1999). In Würmian periglacial areas, the channel contains blocks transported by palaeofloods, which have stopped channel degradation upstream and downstream of the mining sites and now armour the bed.

Changes in conditions along the channel margin can also induce channel changes independently of upstream controls (Fig. 5.1). In the Drôme basin (1640 km^2) in France, a major rockslide in 1442 at km 81 (measured upstream from its confluence with the Rhône) has strongly influenced the upstream channel characteristics. Here, the channel has a gentler slope (0.003 versus 0.005 at the confluence with the Rhône) and a stable single-bed meandering channel. Without the damming effect of the landslide, a steep slope and a braided channel would be expected in this reach. Change in vegetation cover within the floodplain can also induce a channel metamorphosis. Riparian vegetation loss from fire, grazing or mechanical removal can increase bank erosion and favour channel widening and shifting (Orme and Bailey 1970). Conversely, channel narrowing was observed in many rivers in France during the 20th century due to the increase in bank resistance by the establishment of riparian vegetation after abandonment by agriculture or grazing (Liébault and Piégay 2002).

Fluvial systems range in scale from that of the vast Amazon River system (draining nearly 7 million km^2) to small badland basins of a few square metres. Fluvial systems can also be viewed over periods ranging from a few minutes of present-day activity to channel changes of the past century and even to the geological time required for the development of the billion-year-old gold-bearing palaeo-channels of the Witwatersrand conglomerate of South Africa.

To simplify discussion of the complex assemblage of landforms that comprise a fluvial system, its longitudinal dimension is traditionally subdivided into three zones: the sediment source zone, the transport zone and the deposition zone (Schumm 1977). These three subdivisions of the fluvial system may appear artificial because obviously sediments are eroded, transported and stored, in all the zones; nevertheless, within each zone one process is normally dominant through time. However, the sequence of sediment source zone to transport zone, and possibly deposition zone, can be repeated many times along a river with active sediment sources.

Each zone of the fluvial system, as defined above, is an open system. Each has its own set of morphological attributes, which can be related to water discharge and sediment movement.

The hydrosystem concept

As geomorphologists increasingly interact with other environmental scientists, geomorphic processes are considered in relation to biological processes and human actions. Predicting human effects on river systems on a time-scale of multiple decades allows us to evaluate better the societal costs of human actions, to understand tradeoffs between societal uses (e.g. leisure activities, navigation) and natural resources (water, gravel, forest, fish, hydroelectricity) supplied by the river.

The concept of the 'hydrosystem' provides a framework within which to evaluate such interactions (Amoros *et al.* 1988; Amoros and Petts 1993). The hydrosystem can be defined as a three-dimensional system dependent on longitudinal (upstream to downstream), lateral (channel versus margins) and vertical (surficial versus underground) transfers of energy, material and biota (Fig. 5.2). Its integrity depends on the dynamic interactions of hydrological, geomorphological and biological

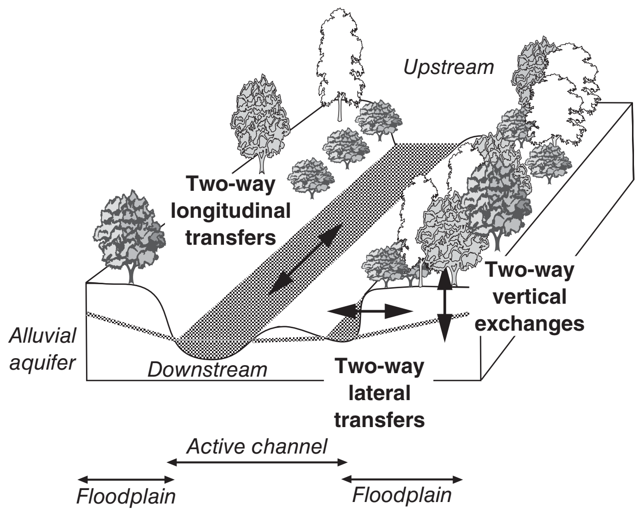


Figure 5.2 The hydrosystem, a complex system with three bidirectional axes: upstream/downstream, channel/margins, surficial/underground environments.

processes acting in these three dimensions over a range of time-scales. The system components are interrelated in the sense that many fluxes may be bidirectional.

The longitudinal dimension is defined by upstream–downstream relationships. For example, alluvial channel form is controlled by the sediment input from upstream and changes in sediment supply can lead to aggradation, no change in bed elevations or incision. In turn, channel changes such as degradation and armouring can influence in-channel features (e.g. pools, riffles, glides) which are critical habitats for fish communities. Downstream factors may also affect upstream factors. A drop in base level (e.g. from sea-level lowering or in-channel gravel mining) can induce regressive erosion upstream, which in turn can expose rock outcrops and undermine check dams, which can become barriers to anadromous fish migration.

In the lateral dimension, the bidirectional links between the main channel and its margins are particularly complex within alluvial corridors. In alluvial valleys, palaeoforms (terraces, alluvial fans, screes) commonly influence channel characteristics. The geological setting influences the slope and width of the valley floor and consequently channel slope and pattern. The lower valley of the Ubaye River, a tributary of the Durance in the southern French Alps, is characterized by an unusual successional pattern (Fig. 5.3). It is braiding across a large valley cut in marl, but becomes progressively more meandering and then straight downstream as it traverses more resistant rocks, becoming narrower, with a higher gradient and coarser bedload (Piégay *et al.* 2000). Channel behaviour controls the floodplain architecture and consequently its biological diversity. A freely meandering river has the capacity to create cut-off channels, within which water bodies support exceptionally diverse ecosystems. Their life span at the aquatic stage depends on their efficiency in terms of sediment trapping (frequency of flooding and critical shear stress) and the upstream characteristics of the basin (sediment supply, flood magnitude and frequency) (Citterio and Piégay 2009).

The third dimension corresponds to vertical interrelationships. For example, channel degradation or aggradation may induce changes in the biological and chemical functioning of the floodplain. Channel degradation, for example, may induce water table decline, terrestrialization of floodplain wetlands, which consequently increases vegetation encroachment and then sediment trapping. Several examples are given in Chapter 10 showing how ecological changes can provide information about the geomorphic adjustment of the river channel.

The hydrosystem concept can be considered as an extension of the fluvial system concept (Schumm 1977), as applied to large rivers with well-developed floodplains. It involves geomorphological parameters but also chemical and biological parameters. Whereas the fluvial system emphasizes temporal and longitudinal dimensions, the hydrosystem concept emphasizes the lateral and vertical dimensions, which are most important on large floodplains and which strongly influence alluvial groundwater storage, ecological richness and regeneration of riparian vegetation. Moreover, large floodplains are often heavily developed, so various uses and engineered structures affect natural processes, channel forms and floodplains.

System theory is increasingly employed within the environmental sciences, with shared vocabulary and explicit integration of physical factors and their relationships in wider systems of biological reactions, as reflected in the serial discontinuity concept (Ward and Stanford 1983), the hierarchy (Frissell *et al.* 1986), the flood-pulse concept (Junk *et al.* 1989) and the riverscape and patch dynamics (Townsend 1989; Thorp *et al.* 2006).

The fluvial anthroposystem concept

Neither the fluvial system nor hydrosystem concepts explicitly consider humans as an element of the system, considering them as external drivers acting on the system, similarly to lithology or climate. Hydrosystems widen the system from physics to biology and chemistry, but not to humans, which are still considered separately. With increased understanding of the strength of human influences on river forms and their processes, it is clear that river evolution also influences social development and human well-being, which depend on services and goods provided by the river (Kondolf and Piégay 2011). A new paradigm is emerging, in which humans are seen as part of river systems and which considers human benefits of river conservation, mitigation, management and restoration. Moreover, 'references' for river restoration need no longer be pristine channel features. Anthropogenic channels may also provide interesting ecosystems, raising new challenges for ecological engineering and fluvial geomorphology.

Thus we come to the scientific debate about what a healthy river is and how river managers can act to improve a river's ecological state. Are anthropogenic rivers fundamentally unhealthy, as implied by many assessment methods under the

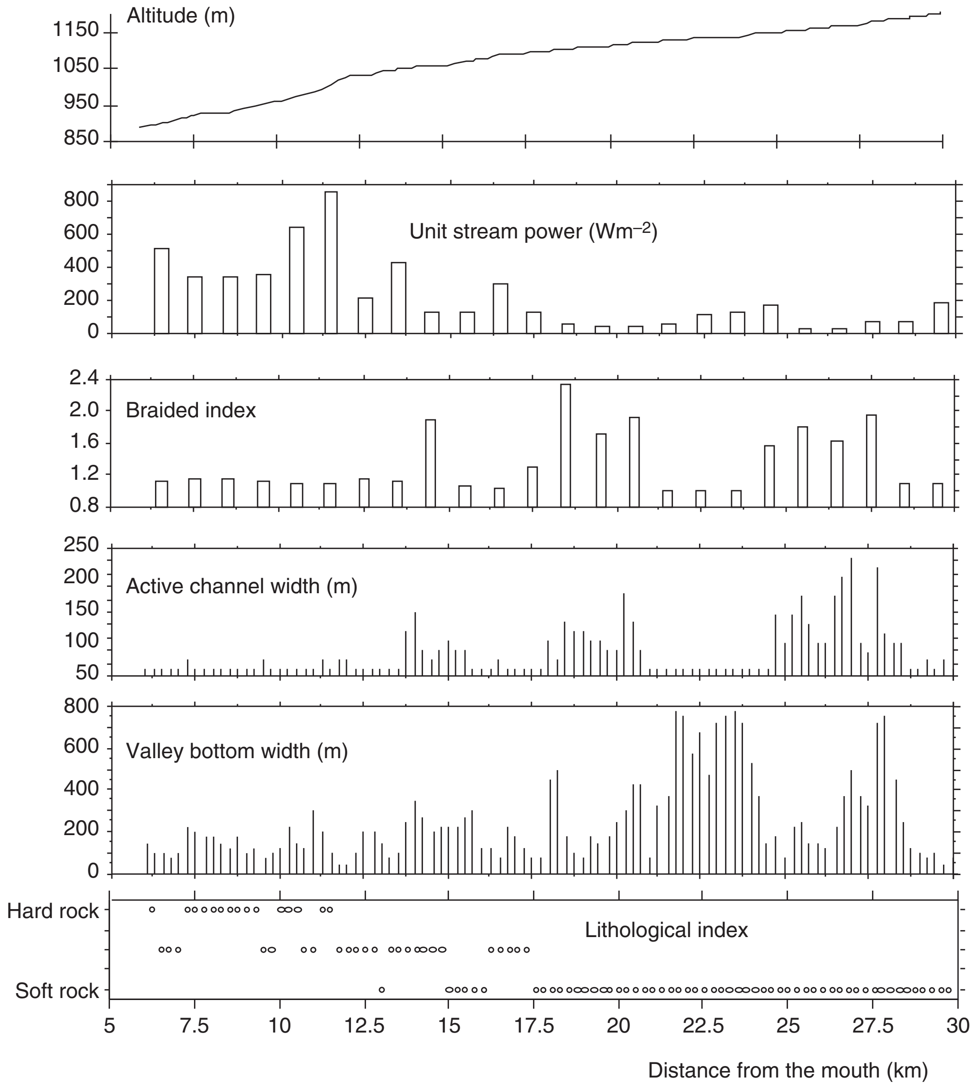


Figure 5.3 The longitudinal continuum observed along the downstream reaches of the Ubaye River (France): a channel slope increase at about 11 km and width decrease associated with modifications of lateral controls (valley narrowing and increase in coarse sediment input downstream). Source: Piegay and Salvador 200a Reproduced with permission of Schweizerbart.

Water Framework Directive? Or can rivers manipulated by anthropic means reach a functional state that is ecologically valuable? The notion of environmental value is increasingly driving how we interpret river condition and social, historical and cultural factors are becoming crucial to understanding how we interpret river changes and their consequences. As an example, the disappearance of braiding in the European Alps has been interpreted by scientists as a human alteration, and loss of habitats associated with actively braiding channels were viewed as an impact. Although this is a valid concern, it is important to recognize that some of the braiding activity was the result of deforestation and overgrazing during the 19th century and it was the reduction in sediment delivery resulting from a spontaneous afforestation that led to narrowing of channels and, in many cases, consequent expansion of the riparian corridor, a fantastic 'renaturation' of the mountain area due to the decline in human pressure on the landscape.

After centuries of human policy promoting works to control and suppress river processes to favour human development, there is increasing interest in adopting river management approaches that balance development with ecological values, allowing natural river processes to function unhindered as much as possible, while recognizing the inevitability of human alteration to the formerly pristine system. This implies that restoration should be guided by a functional paradigm that views a healthy river as functioning well and being complex in terms of habitats, rather than attempting to emulate a past reference condition. In this context, river evolution can be conditioned by both natural and anthropogenic drivers.

5.2 Components of the fluvial system

Scales of analysis and the range of influencing factors

The fluvial system and its components can be considered at different spatial scales and in greater or lesser detail depending upon the objective of the observer. The river basin is a critical component as it provides floods and sediments conditioning channel forms, so that its evolution is of interest for exploring factors controlling channel changes. At a reach scale, the channel pattern reveals such river history. At a finer spatial scale, those of a channel feature, such as a single meander bends, flow hydraulics, sediment transport and rate of meander migration can be measured. Within the channel feature itself, grain size patch is also a valuable spatial scale providing information on sediment sources or sediment loads.

The fluvial system is characterized by an asymmetry of controls in the sense that the broader scale levels influence the smaller scale levels (e.g. influence of basin scale on reach scales), whereas the inverse is rarely true. In the same way, changes affecting a given reach influence the structure and functioning of the sediment facies and the vegetation units. For example, a dam may provoke downstream incision and bed coarsening (boulder exhumation) and simplification of the channel pattern. If the broader levels are not considered, the ecologist may be unable to explain fish abundance and diversity at a reach scale relative to those elsewhere, or to design sustainable restoration or mitigation actions.

Various components of the fluvial system can be investigated at different scales, but no component should be totally ignored, because hydrology, hydraulics, geology and geomorphology interact at all scales as human pressures. This emphasizes that the entire fluvial system should not be ignored, even when only a small part of it is under investigation.

Non-linear temporal trajectory of fluvial systems

The fluvial system provided a clear framework to theorize the temporal evolution of rivers, considering the complex time pattern and developing conceptual basis: adjustment, resilience, threshold and sensitivity, but also lag time.

Two major time-scales are then distinguished: temporary channel changes linked to a punctual event, which strongly modify the channel, such as a large flood, versus long-term, irreversible changes. The concept of 'dynamic equilibrium' considers the channel to be metastable around an average condition, fluctuating in response to the magnitude and frequency of floods (Hack 1960). This dynamic equilibrium is broken when a river adjusts to new control conditions and shifts to a new equilibrium position. River systems may be more or less resistant or sensitive to changes. A bedrock channel is evidently very resistant whereas an alluvial system draining sandy environments with a low vegetation cover will be much more sensitive to a modification of its hydrology or of sediment delivery. Moreover, changes in controlling factors have different effects on channel geometry according to their own pattern of change and the distance between each other. The concept of 'geomorphic thresholds' is relevant (Schumm 1973). Channel changes may occur slowly and in a continuous manner up to a threshold, a critical time step above which change is observed. Moreover, channel changes tend to lag behind changes in the controlling factors. This lag time depends not only on the processes driving the cascade of changes downstream, but also the channel's sensitivity to changes. Following the fluvial anthroposystem theory, with human pressures ubiquitous for many centuries, rivers are reacting to multiple human pressures acting at different time-scales and different locations within the basin, inducing different lag times for channel responses and producing different combinations of responses (e.g. channel narrowing versus widening, incision versus aggradation, coarsening versus fining, increase versus decrease in lateral shifting). Figure 5.4 illustrates river trajectories as affected by control factors operating on different temporal patterns. The relationships displayed in Fig. 5.4 are straight forward. They demonstrate that, because of the number of variables acting, the fluvial system has a complex history, as it adjusts to climatic changes and human influences through time. In addition, at any one time, the range of geology, relief and climate guarantees that a great range of morphologic characteristics can exist among drainage basins.

When multiple control factors operate, their effects can interact, in some cases one factor counteracting another, and in other cases combining to produce unexpected critical changes. This idea of complex non-linear fluvial system responses is well illustrated by the emerging concept of trajectory, a more complex concept of adjustment, which reconsiders notions of reversibility, resilience and recovery, especially in the context of restoration in response to human alteration (Brierley *et al.* 2008), and follows the idea of viewing rivers as non-linear dynamic systems where forms are controlled by chaos and self-organization (Phillips 1996). If rivers are continuously changing in response to multiple pressures, then any return to a previous state is difficult and 'restoration' to a historical reference condition impossible. The concepts of cyclicity and resilience, expressed as the time needed for a system to return to its pre-disrupted conditions, may conflict the trajectory model, being evident for perturbation (linked to a punctual event that occurs at a given frequency) but not for alteration (linked to an unexpected factor that occurs not cyclically and with a magnitude that is also variable). These concepts of disturbance and recovery are critical for river management because they form the implicit basis for diagnostic studies of causal factors for management problems. These concepts can also provide information on how the river works and how sensitive it is to actions, and provide expert knowledge for designing actions to manage sustainably, to improve or to repair.

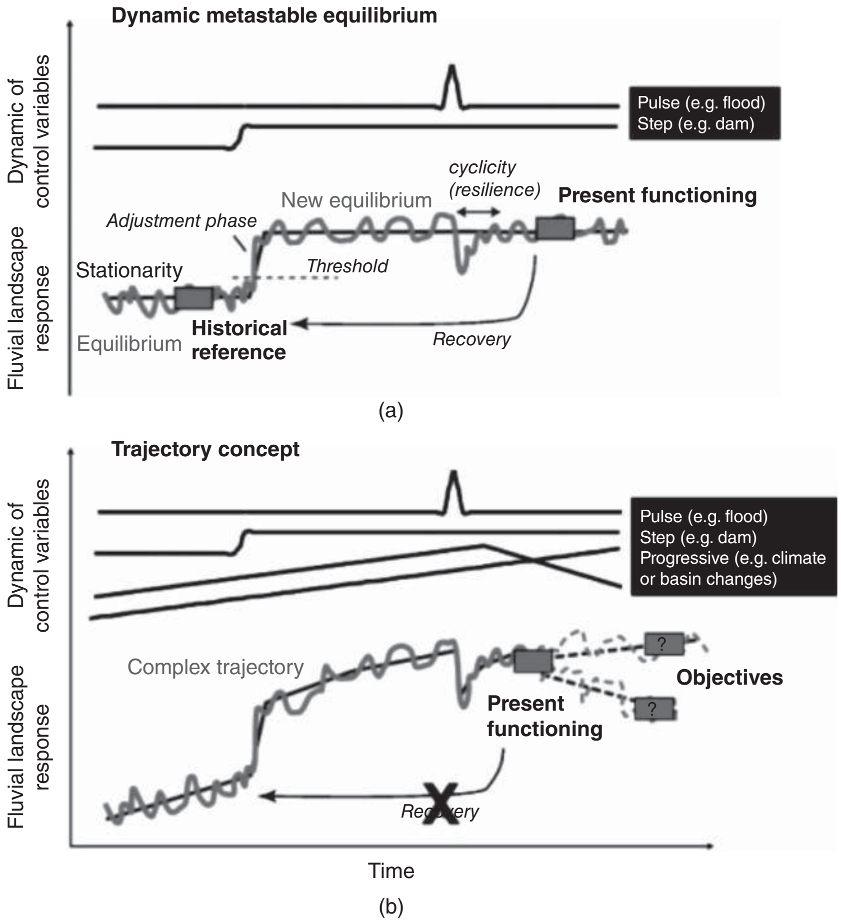


Figure 5.4 Example of temporal trajectory followed by a fluvial landscape due to its adjustment to a complex spatio-temporal framework of control factors according to the dynamic equilibrium and trajectory concepts. Source: Dufour and Piégay, 2009. Reproduced with permission of Wiley.

5.3 Fluvial system, a conceptual tool for geomorphologists

Partial versus total system approach

There is more than one approach to the fluvial system. One may be ambitious and attempt a total-system analysis integrating information on all aspects of the fluvial system, but usually there will be insufficient information to permit such an approach. Rather, one may choose to investigate only the source zone or a channel reach. This reduced partial-system analysis is usually all that can be attempted, but its importance lies in the value of viewing a limited problem or limited study area in a broader perspective.

If geomorphological studies are characterized by spatial limits, they have also temporal limits. In this context, a system approach is always partial because time in the fluvial system is not bounded like a basin but may change at the seasonal, decadal, century or Holocene scales. Geomorphic systems can be studied at different time-scales depending on the study objectives, which are to explain present geomorphic features or their sensitivity to changes in runoff and sediment yield, but the observer must have always in mind that their observations apply in a given temporal setting. On gravel-bed rivers, for example, the observed channel width integrates effects of the last large flood (vegetation scouring versus encroaching) and longer term effects in terms of sediment delivery or floodplain land-use changes, which influence its resistance and local shear stress. Depending on when the channel width is observed, it may be experiencing different trends of short-term evolution (narrowing versus widening).

The fluvial system, a concept for structuring hypothesis

The 'fluvial system' can be seen as a conceptual model developed by the researchers based on their results. Complexity is added to the original simple model through regional studies, which show the importance of effects such as riparian vegetation and geomorphic facies and the cascading effects of geomorphic changes on living communities and human uses (Bravard *et al.* 1997; Pont *et al.* 2009).

Once the conceptual model has been defined, it can be used as a tool to focus efforts early in research process. It is then a basis to formulate hypotheses in a deductive approach, allowing the researcher to build a preliminary, rough architecture of the studied component to test the potential factors controlling it, its sensitivity to changes, the acting range of its processes and forms, the geomorphic thresholds. Thus, the fluvial system provides a simple framework into which complexities of the specific river can be placed in contrast and within which questions can be posed, such as the potential effects of changes in peak flows or sediment load, or when currently occurring adjustment is likely to be finished.

Such a conceptual approach is now applied within the anthroposystem framework for restoring rivers considering both natural parameters (physics, ecology, chemistry) and socio-economic drivers that influence the natural parameters and condition actions (Fig. 5.5, from Jacobson and Berkley 2011). As explained by these authors, such conceptual models are useful in adaptive management projects to visualize and share understanding of how a river works and what can be the potential cascading consequences of restored actions discussed. It is then a working tool that both allows scientists from different disciplines to build a common conceptual model and also aids in public education and communication.

The comparative space–time framework

The fluvial system concept can serve first to integrate case studies in a broader spatial and temporal scale context, considering upstream influences on channel and long-term trends. Many monographic studies, such as those of Bravard (1987) on the upper Rhône River (France) and Agnelli *et al.* (1992) on the Arno River (Italy), have been carried out in this perspective and brought many useful elements to underline the complex history of channels. Although a single case study can facilitate the

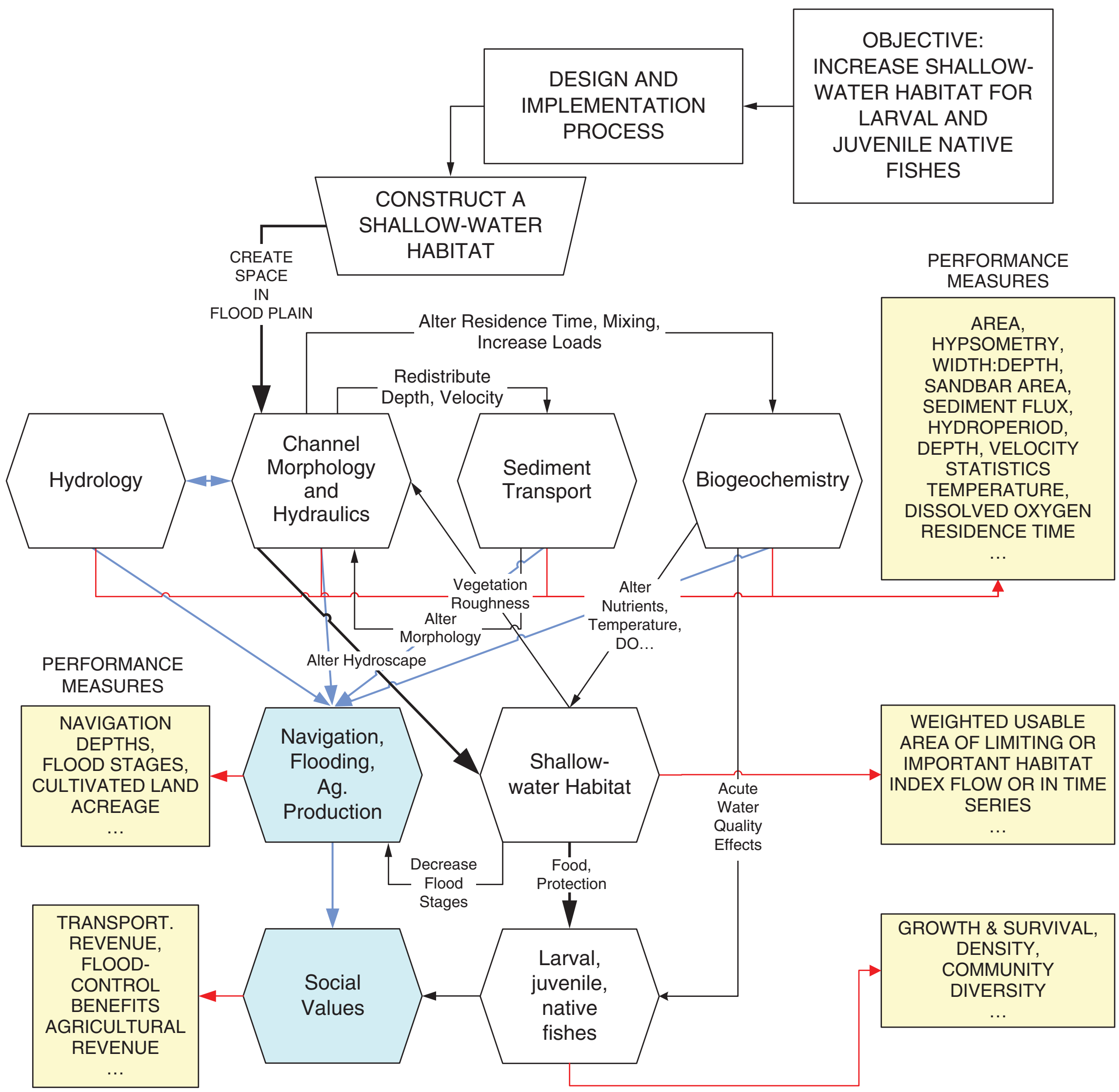


Figure 5.5 Conceptual model applied to shallow water habitat restoration on the Lower Missouri River. Shallow water habitat construction is intended to restore rearing habitat for larval and juvenile pallid sturgeon. From Jacobson and Berkley (2011, Fig. 6, p. 23). Source: Jacobson and Berkley, 2011. Reproduced with permission of AGU.

understanding of channel change and structural complexity, it is often risky to generalize the results to provide a reproducible model of how the river functions and is sensitive to given acting forces. In such a context, the fluvial system concept facilitates the development of comparative studies for generalization purposes and theory.

The analysis of the fluvial system can then be based on comparisons of many spatial units corresponding to *components* of the system. As noted above, the fluvial system is composed of different open and interacting *components* (e.g. geomorphic units) which are nested (drainage pattern > river reaches > channel features > sedimentary patches) and described by *attributes* (e.g. a channel reach can be defined by its geometry, water and sediment processes, morphological changes). Each component can be compared with others at a single scale by comparing their attributes or the study can focus on the interactions between the nested components. Two conceptual approaches can be then distinguished. By comparing a set of components of the fluvial system at any spatial scale or within a temporal perspective, *similarity analyses* distinguish them according to attributes, order them according to key geomorphic questions (such as the stage of evolution and specific process response) and then build conceptual models. *Connectivity analysis* can aid in understanding the cascading factors that control the changes of nested components and in building causal and chronosequential models.

Thus, s*imilarity analysis* and *connectivity analysis* are two ways to study the fluvial system in a comparative manner, one focusing on single temporal and spatial scale components, the other on the links between components of different scales. The size and heterogeneity of the study area, the question posed and the causes of changes should determine which approach (or combination of the two) is chosen. Within a *connectivity approach*, *similarity analysis* can be carried out at each scale level if a set of components is studied.

In this context, the ways of approaching a river can be summarized by a set of 3D diagrams (Fig. 5.6), each axis being respectively the spatial scale level (e.g. in-channel feature unit, channel reach unit, floodplain unit, basin unit), the time-scale level (season, year, decade, century) and the number of spatial units or *components* being considered. The basic approach focuses on a single spatial unit observed at a single scale level and without temporal perspective (upper left diagram). This is the first level of a case study, the description of the geomorphic characters of the spatial unit, which usually is augmented by studying its sub-units in an integrated perspective and its changes over time. The ultimate level of the case study approach considers all the characters of the spatial unit: its inner complexity and the relationships between its sub-units (*connectivity analysis*) through time (lowest left diagram). A similar approach can be taken in a comparative perspective (right side of diagram). Rather than describing a single spatial unit, many are described simultaneously to identify differences among them or to order them on a longitudinal or a temporal gradient (*similarity analysis*). The second level of comparative studies, which is one of the most developed in the geomorphic literature, is to compare nested spatial units, typically a channel reach and its basin. This is the basis of hydraulic geometry analysis (Hey 1978 , Ferguson 1986) or allometric studies, which are based on empirical power functions, relating basin size, usually basin area, to channel geometry (e.g. width, depth, cross-sectional area, channel length, area of alluvial fans) (Church and Mark 1980). Comparative studies can also consider the temporal trend of each spatial component to consider differences in adjustment rather than differences in structure. These different comparative studies can ultimately combine both *similarity* and *connectivity* approaches.

Similarity analysis

Similarity analysis focuses on a set of landforms at a single spatial scale level for which the comparison of the attributes allows groups to be identified and ordered. This approach is commonly used in partial-system analyses. Similarities or dissimilarities can be assessed at various spatial scale levels: e.g. a set of channel reaches to study similarities within a set of bars, a set of basins or of reaches within a basin to study similarities within a set of channel cross-sections. This analysis is mostly synchronous in the sense given by Amoros and Bravard (1985), involving essentially simultaneous measurements at numerous sites, results of which can be analysed statistically.

Similarity analysis can be also conducted in a temporal perspective. *Diachronic* or *retrospective analysis* (Amoros and Bravard 1985), involving assessment of changes of a single components over time, using historical sources, is important in this approach. According to the intensity, spatial extent and chronology of these changes, it is then possible to evaluate the causes of changes of the dependent components and their respective importance and to assess their present and future effects.

Two tools or approaches that are particularly useful in such investigations are *location for time substitution* to develop evolutionary models of landform change and *location for condition evaluation* for assessing landform sensitivity or resistance (Schumm 1991) (Fig. 5.7a and b).

The *location for time substitution* (LTS) is a well-known tool to geomorphologists and it is often referred to as the 'ergodic' method or 'space for time substitution', but following Paine (1985) and Schumm (1991), it will be referred to here as location for time substitution. This involves the selection of a sample of components that can be arranged in a sequence that shows change through time.

The second technique can be used to identify sensitive components in what can be termed the *location for condition evaluation* (LCE). This involves measuring the characteristics of relatively stable and unstable components. A comparison permits the identification of critical threshold conditions and sensitive components. Using such a technique, we assume that each case is in dynamic equilibrium with its controlling parameters,

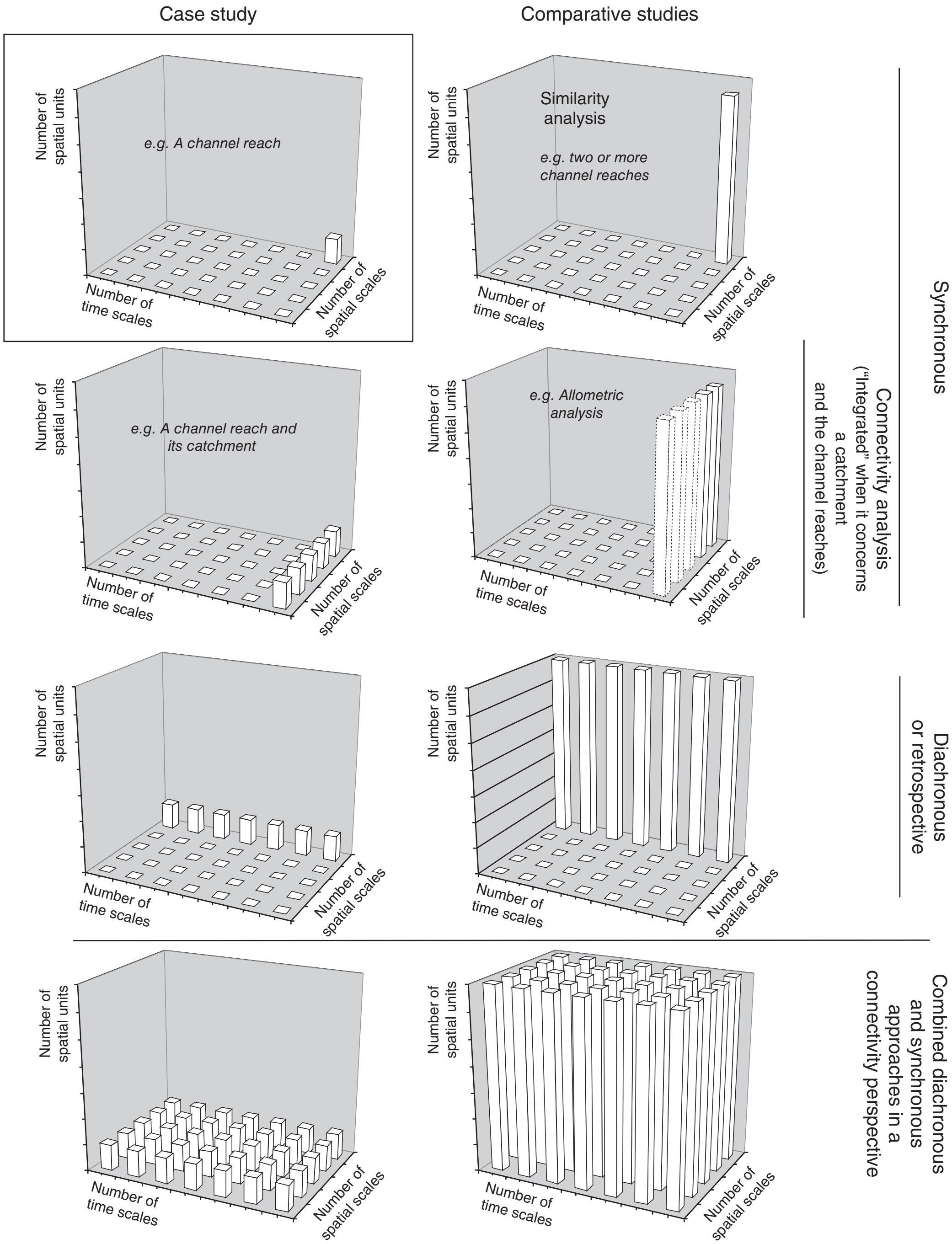


Figure 5.6 Schematic 3D models to summarize how fluvial systems can be studied based on time-scale, spatial scale and number of spatial units or components considered. A number of different approaches are then possible: monographic or comparative, synchronous or diachronous, based on *similarity* or *connectivity* assessment. Source: Pont *et al*, 2009. Reproduced with permission of Springer.

which means that threshold conditions depend on the intrinsic characteristics of the system (e.g. geological or climatic setting).

One of the best known LTS analyses has been used to show incised-channel change with time. A series of cross-sections surveyed along a channel that has incised illustrate an evolutionary model of channel adjustment resulting of natural or human-induced changes (e.g. channelization) (Fig. 5.8). Although this model would not be expected to apply to all streams, experimental studies support the model's sequence for many incision channels (Schumm *et al.* 1987) and the model has been further developed for a range of applications (Simon 1989). The model presented in Fig. 5.8 was developed for incised channels in northern Mississippi and it has both academic and practical value because it permits the estimation of sediment

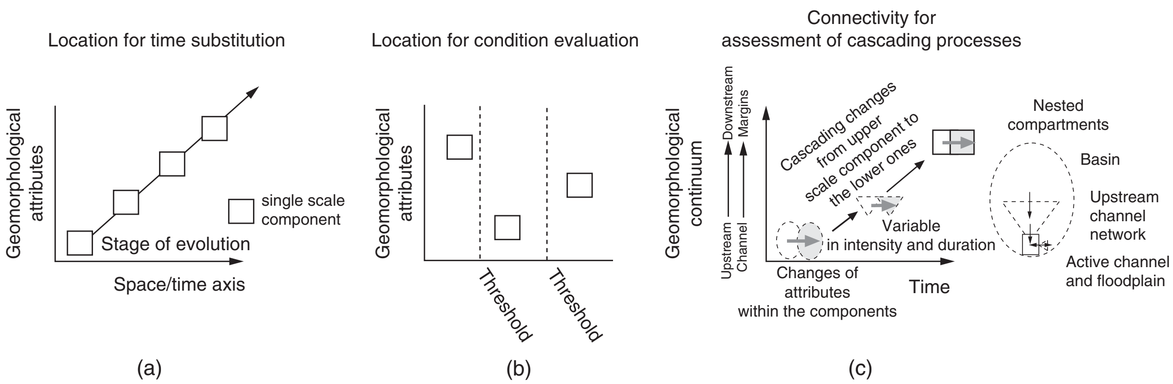


Figure 5.7 Summary of the different conceptual models used in fluvial system analysis: (a) location for time substitution; (b) location for condition evaluation; (c) connectivity model.

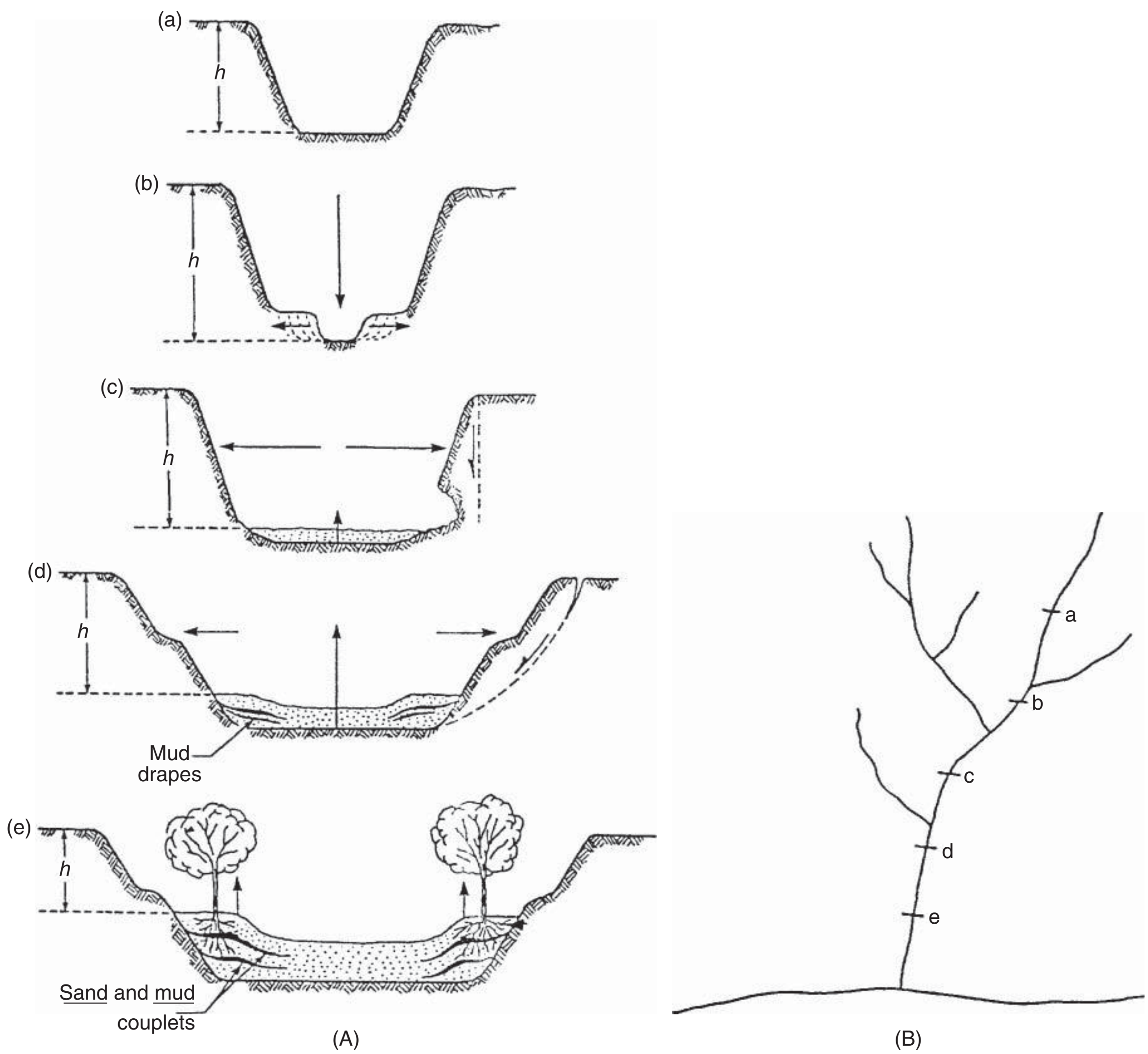


Figure 5.8 (A) Evolutionary model following a location for time substitution analysis (LTS) of incised channel from initial incision (a, b) and widening (c, d) to aggradation (d, e) and eventual stability (e). Modified from Schumm, *et al*. (1984). (B) Measurements at sites a–e provide information for the development of a model of channelized stream (incised channel) evolution.

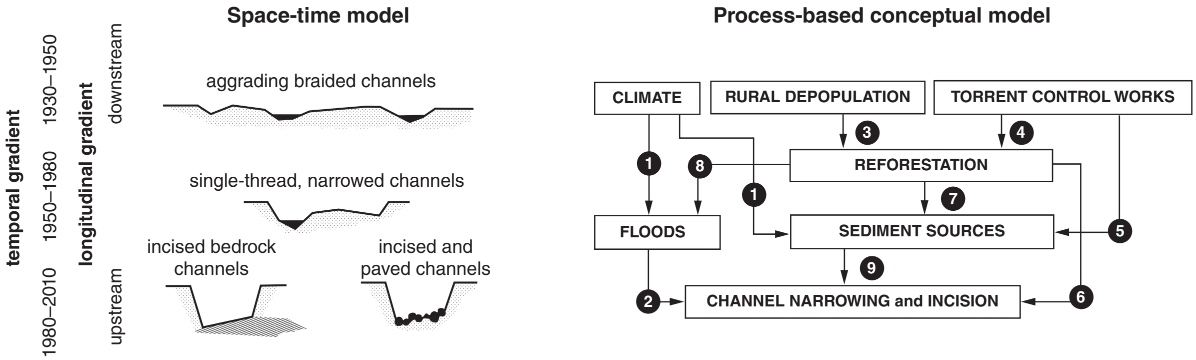


Figure 5.9 Location–time and causal, process-based models of adjustments between geomorphic channel stages observed in the basin. The process-based model shows cascading responses among different factors explaining the observed adjustments. 1, Decreasing frequency and/or intensity of extreme rainfall events; 2, decreasing frequency and/or intensity of channel-forming floods; 3, spontaneous afforestation of hillslopes and valley floors due to the abandonment of agricultural uses; 4, reforestation for soil conservation and construction of check-dams in steep headwater reaches; 6, Increasing the bank roughness and stability by riparian vegetation and decreasing peak flows; 9, decreasing the sediment supply. From Pont *et al.* (2009).

production and agricultural land loss and the identification of channel reaches that require controls (Schumm *et al.* 1984). The location for time substitution can be an effective means of developing a model of evolving landforms. Research carried out in the European Alps also showed different adjustment patterns to channel incision (Liébault and Piégay 2002). Whereas the Schumm model predicted channel widening in the loess landscapes of the Mississippi delta, the Alpine model involves channel narrowing following incision (Fig. 5.9) (Pont *et al.* 2009).

Research in central southern England (Gregory *et al.* 1992), Zimbabwe (Whitlow and Gregory 1989) and Arizona (Chin and Gregory 2001) used a LTS framework and a downstream hydraulic geometry analysis based on the empirical relationship between channel cross-sectional area at bankfull and the basin area. Urbanized basins were characterized by increased flood frequency and consequent channel degradation and widening. As a result, the urbanized basin channel deviated from the general relationship by being wider and deeper than what the model predicts. Data collected in Fountain Hills basins (Arizona) showed the expected downstream increase in channel width, depth and capacity with drainage area whereas the data collected in reaches disrupted by urbanization yield channel widths up to two times wider than expected from undisturbed longitudinal patterns alone.

When using LTS it is then important to compare features produced by the same processes that are operating under the same physical conditions. For example, the evolution of an incised channel in alluvium can be determined by surveying cross-sections at several locations where the channel is in alluvium, but one cannot combine data or compare channels in weak alluvium with channels in resistant alluvium or bedrock and expect to find meaningful results. Therefore, if one is asked to evaluate the stability of a site, it is wise to search for similar site conditions within the same general area and with a comparable geological setting and to use these to aid in the specific site evaluation. Observed sequences reflect exclusively the temporal evolution step-by-step, which means that we assume that the studied features are in desequilibrium with their controlling factors.

Location for condition evaluation has been used to identify sensitive valley floors that are likely to gully in Colorado (Fig. 5.10) and New Mexico (Patton and Schumm 1975; Wells *et al.* 1983), river reaches that are susceptible to a pattern change from straight to meandering to braided (Fig. 5.11) and alluvial fans that are susceptible to fan-head incision (Schumm *et al.* 1987) and thresholds of hillslope stability (Carson 1975). This approach is similar to the location for time substitution, as described above, except that it is the present conditions rather than an evolutionary model that need to be evaluated.

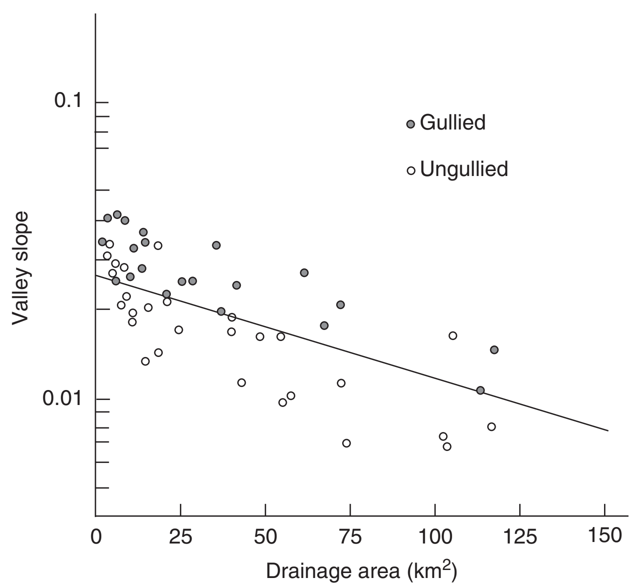


Figure 5.10 Location for condition substitution permits identification of threshold valley floor slope at which gullies form. Source: Patton and Schumm, 1975. Reproduced with permission of GSA.

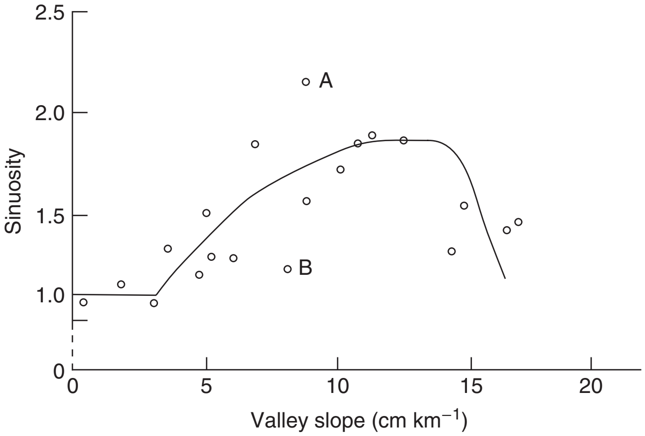


Figure 5.11 Location for condition substitution of valley floor slope at which the 1890 Mississippi River channel pattern changed from low sinuosity to meandering and from meandering to a transition meandering–braided pattern (high to low sinuosity) as the valley slope varied. (A) river reach where cut-offs will occur soon, and (B) river reach adjusting for recent cut-offs and sinuosity. Adapted from Schumm *et al*, 1972.

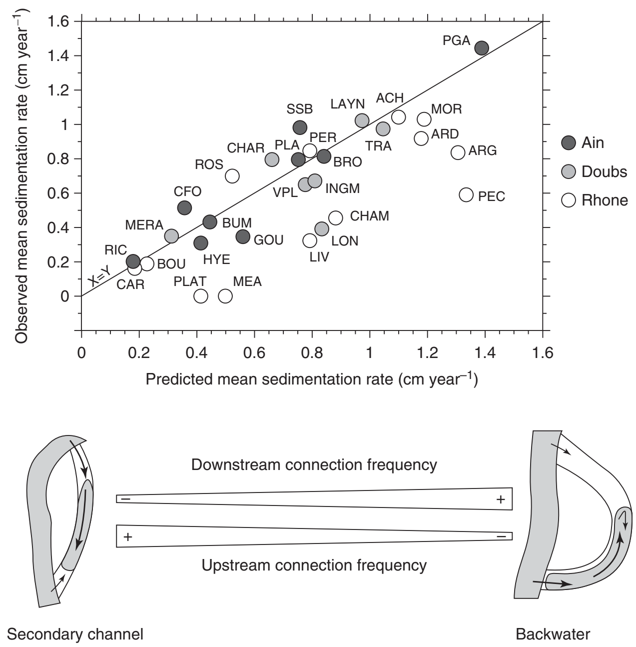


Figure 5.12 Observed versus predicted sedimentation rate from multiple regression model based on frequency of upstream and downstream overbank flow frequency. The model is based on samples from the Ain and the Doubs and is used to predict the sedimentation rate of the Rhône former channel lakes. Source: Citterio and Piégay, 2009. Reproduced with permission of Wiley.

In each of these cases, data were collected at a number of locations and a relation was developed to identify future or threshold conditions. For example, the slope of the line in Fig. 5.10 identifies a valley floor slope at a given drainage area at which gullies are likely to form. When a relation such as that in Fig. 5.10 is developed between drainage area and alluvial fan slope, alluvial fans that are susceptible to fanhead trenching can be identified. The curve of Fig. 5.11, when developed for a specific river, can be used to identify when a river pattern is susceptible to change from meandering to braided and vice versa. In addition, the vertical position of a point on the plot is an indication of future change. For example, the point with the highest sinuosity (A) represents a river reach where cut-offs will occur, whereas the point that plots very low (B) is adjusting for previous cut-offs and sinuosity will be increased by meander growth.

The floodplain of the Ain, Doubs and Rhône rivers in eastern France has numerous cut-off channels, which range widely in habitat types as they evolve from fully aquatic to terrestrial, as they silt up through time. This range of habitat types results in high biodiversity. LCE analysis has been conducted on a set of 39 cut-off channels in order to understand their silting dynamics and assessed their sensitivity to terrestrialization (Citterio and Piégay 2009). The conventional model of sedimentation rate, decreasing as a function of time, such as those established by Hooke (1995), has not been observed in these cut-off channels, the youngest forms (20 years old) having sometimes thicker overbank sediment than others that are 65–80 years old. LCE analysis showed that fine sedimentation in cut-off channels is controlled by overflow frequency from both upstream and downstream entrances which are linked to channel planform types when cut-off occurred (Fig. 5.12). When the upstream overflow frequency is high, cut-off channels undergo scouring, preventing strong sedimentation. Conversely, when downstream overflows are more frequent than upstream overflows, backwater events exacerbate sediment entrances and deposits. The meander cut-off channels are mainly flooded by backwaters and consequently experience high deposition rates, whereas braided cut-off channels function as secondary channels during floods and are less susceptible to fine sedimentation due to high flow velocities. Owing to channel metamorphosis during the 20th century, braided cut-off channels are often older than meander ones, explaining the complex pattern of sedimentation rates in this regional setting.

Along the Ain River, vegetation encroachment and channel narrowing in the 20th century were initially attributed to decreased peak flows due to an upstream dam built in 1968. However, studies of the chronology of vegetation encroachment on the Ain and other Rhône River tributaries (Fig. 5.13) showed that the encroachment could not be explained primarily by dam-induced flow changes (Piégay *et al.* 2003). Vegetation encroachment began on the Ain before the dam and was observed on other rivers whose peak flows were not disrupted by dams, evidently affected by land-use changes on the floodplain and in the basin. Two main types of vegetation encroachment were observed: (i) early in the 20th century, vegetation encroached along braided mountain reaches (such as along the Ubaye River in the 1920s) in response to decreased bedload supply from afforestation of the basin and installation of check-dams from 1880 to 1910; (ii) vegetation encroached from 1945 to 1970 along the Ain and other rivers in the region (e.g. the Eygues, Roubion, Drôme, Ouvèze, Loire

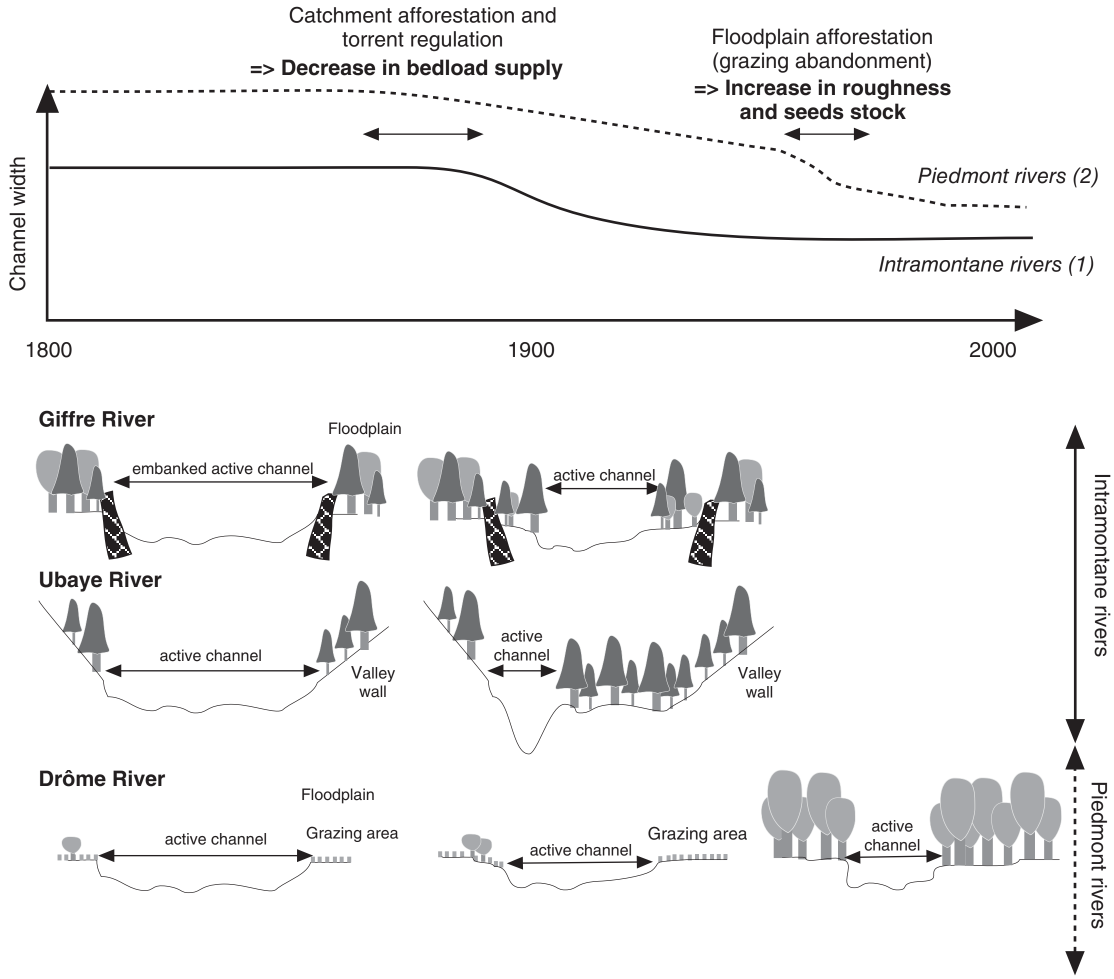


Figure 5.13 Conceptual model summarizing causes and chronology of channel narrowing affecting the alpine and piedmont tributaries of the Rhône River (France) during the contemporary period.

and Allier Rivers), whether influenced by dams or not. This encroachment occurred as the floodplain was abandoned, pastures were replaced by forest and trees colonized gravel bars. In this study, a historical analysis conducted on the Ain showed that the vegetation encroachment preceded the dam. This was then confirmed for other rivers in the region. The multiple case studies were the basis for an LTS analysis, from which a conceptual model of channel changes over the 20th century was developed. The chronology of changes was key to understanding the causal relations. Mountain reaches located close to the sediment sources were the first to show vegetation encroachment, due to decrease in sediment delivery, whereas piedmont reaches downstream experienced encroachment later, probably reflecting both floodplain land-use change and lately decreased sediment input.

Geomorphologists collect data at many locations to develop evolutionary models (LTS) or to determine the sensitivity of landforms (LCE), for practical purposes of prediction, and for environmental reconstruction. Of even greater value, both techniques require that the investigator back away from a single site and look at many sites, which provides the ‘big picture’ and a basis for generalization.

Connectivity analysis

Connectivity analysis focuses on the links between nested components of the fluvial system (e.g. basin and channels, channel reach and former channels) to evaluate better the sensitivity of the lower, dependent components to changes in processes in upper component (Fig. 5.7c). In this context, it is not a comparative analysis of attributes of single-scaled components, but it is a comparative analysis of changes affecting different sets of single-scaled components from which causes (agents, chronology) can be determined. *Connectivity analysis* can also be termed integrated analysis when it concerns links between a basin and the channel reach (Van Beek 1981; Kirby and White 1994). *Connectivity analysis* means that elements are interacting in a bounded physical system, here the basin. It focuses upon the relationships between *components* (e.g. relationships between sub-basins, between river reaches, between a basin and the river channel or between a river channel and its floodplain) integrated within the basin.

Diachronic analysis can identify different adjustments occurring in various components of the system and the timing of those adjustments, especially in relation to changes in land use or other independent basin variables. The combination of these analyses can provide insights relevant to management questions such as causes of coastal erosion when riverine sediment supply has been reduced by dams or in-channel mining.

Connectivity analysis is based on a 'hydrosystem' framework, which assumes that geomorphological attributes of a component result from multiple adjustments, which have cascading effects on the other attributes (biological ones mostly). The aim is to highlight the cascading causal factors of observed changes and to estimate the relaxation time (Fig. 5.7c).

Such models can be time oriented (Fig. 5.14) or component oriented (Fig. 5.15). In the first kind of models, the studied component is the end of a nested system with higher hierarchical levels. When the changes affecting the different hierarchical levels are studied and dated, it is possible to plot them on a temporal axis and identify higher scale changes that explain those observed at a lower scale. The common example concerns the links between a basin and its channel network, but links between a channel reach and its former channels can also be analysed, leading to conceptual models of the effects of one level on others, evaluating the intensity and duration of the propagation of changes downstream or from the channel to its margins. In component-oriented models, the temporal scale and relaxation time scale are less established, but the cascading changes of attributes from one component to another are more clearly modelled.

The East Fork of Pine Creek, Idaho, provides a good example of time-oriented *connectivity analysis*. Kondolf *et al.* (2002) documented channel widening in the 20th century and identified two potential causal factors: (i) a bedload supply increase from the sub-basins caused by mining activities and mining waste inputs and (ii) an increase in bank sensitivity to erosion from grazing and logging on the floodplain. Detailed study of the chronology of the geomorphological phenomena and their potential causes indicated that the first factor predominated (Fig. 5.14).

Good examples of *connectivity analysis* with component-oriented modelling were given by Bravard *et al.* (1997). They described the general trends in river incision in France during the 20th century, underlined the causes and geomorphological consequences and effects of incision on ecosystems of the alluvial plains, such as riparian vegetation, aquatic vegetation of former channels, benthic and hyporheic macroinvertebrate communities and fish assemblages. Conceptual models of cascading factors from geomorphological components to biological components show how vertical channel changes (e.g. aggradation, incision) affect interactions between the channel and former channels, notably rates of fine sediment deposition and rates of vegetation succession (Fig. 5.15).

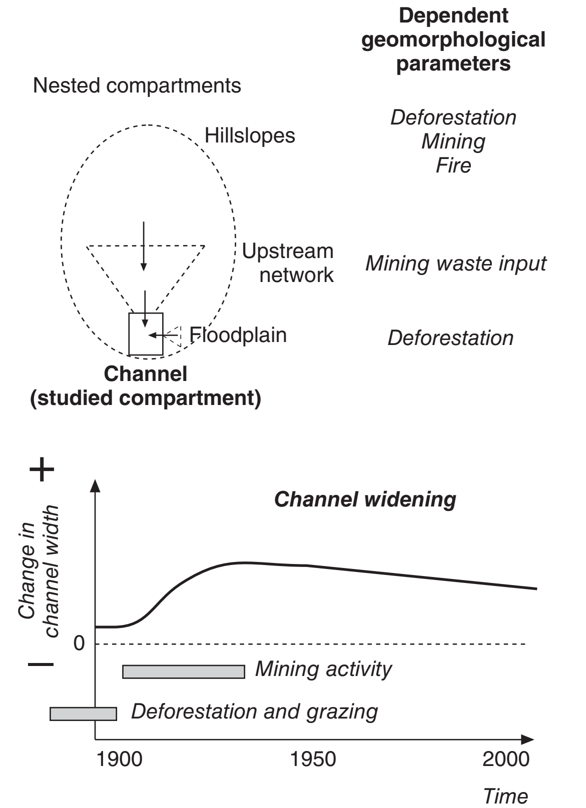


Figure 5.14 Conceptual models of time-oriented connectivity analysis: effects of basin land use changes on downstream channel morphology (East Fork Pine Creek, Idaho).

Quantitative versus qualitative analysis

The systems approach is flexible, in that it can be fully qualitative but also very quantitative, supported by experiments and simulations or intensive characterization of spatial and/or temporal frameworks. It can be developed with increasing precision from expertise to detailed scientific analysis of each of its components and it can be adapted to the management needs of a particular river. Fully qualitative approaches (e.g. geomorphological expertise) are popular in river management to assess different engineering options.

One can use the system concept to pose preliminary hypotheses, then as a framework within which to combine other geomorphic tools. Such holistic approaches are best used in conjunction with reductionist approaches, the first providing an understanding of the river functioning at a large spatial and temporal scale while the second can test hypothesized linkages and simulate processes and changes (Richards 1996).

Experimental (flume) studies have been widely used to validate preliminary hypotheses posed by a larger systemic approach (Schumm *et al.* 1987). The approaches are complementary, as field observations can reveal the complexity of the systems without clearly distinguishing the respective importance of controlled factors, which is more accurately done by experiment.

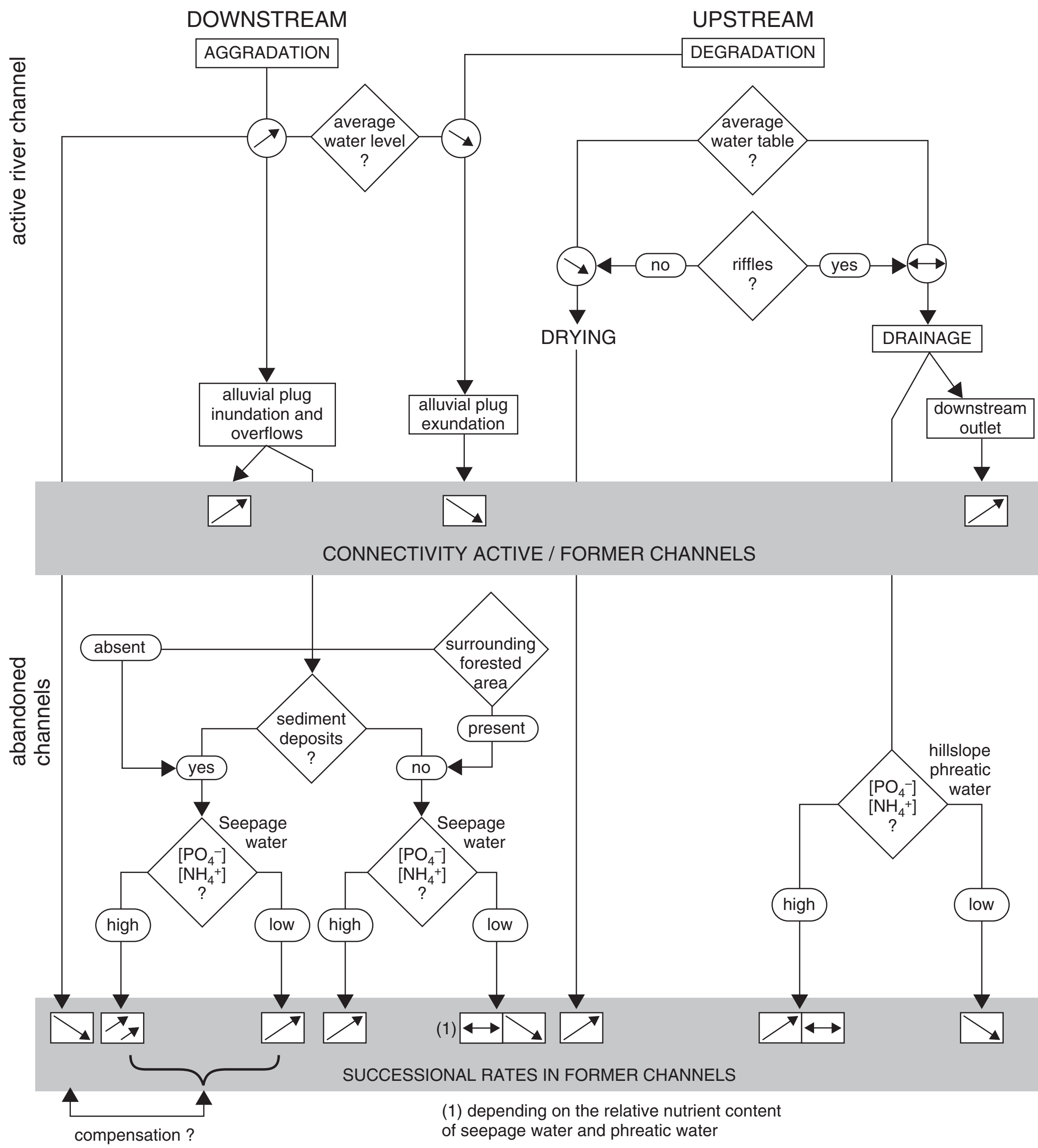


Figure 5.15 Conceptual model predicting effects of vertical main channel changes on vegetation successional rates in former channels (Bravard *et al.* 1997). Two components are distinguished, the main channel (the upper scale level) and the former channels (the lower scale level) and also the changes of their attributes with channel degradation. The long arrows express the cascading effects from one attribute to the other and the arrows in squares express the sense of these effects (increase, decrease or constant).

Empirical approaches with large sample sizes can help calibrate models based on physical laws and identify their boundary conditions and the extent of their applicability. This can be done with comparative studies to identify thresholds or correlations between attributes of components in *similarity* or *connectivity approaches*. Allometric analysis is a well-known quantitative approach based on the fluvial systemic, which facilitated major developments in geomorphology (Church and Mark 1980). The fluvial system concept also underlined efforts to establish discriminate models of fluvial pattern (Bridge 1993) and regression models linking discharge and channel forms (Hey and Thorne 1986). All these empirical models are based on large samples of spatial objects, each one being a binomial of nested components basin–channel, to establish similarities.

Moreover, *similarity analyses* facilitate the development of field experimental studies to test hypotheses and identify threshold conditions. Comparative analysis (paired or multi-basin approach) can then be used to evaluate the effects of human actions on the natural environment, as done by Trimble (1997) at a short time-scale and by Brooks *et al.* (2003) at a longer time-scale concerning riparian vegetation effects on channel geometry. These approaches can be used in river restoration projects for which experimentation in natural conditions is necessary to improve proposed mitigation measures. Henry and Amoros (1995) proposed to compare a restored reach (geometric modifications of cut-off channels and their hydrological connections with the main channel) with a control reach unaffected by restoration works but whose its functioning is similar to it, to distinguish effects of the intervention from other system-wide influences (Fig. 5.16). Such approaches have been used to assess restoration project effects on invertebrate populations (Friberg *et al.* 1994) and fish populations (Shields *et al.* 1997) sensitive to habitat changes. Using *similarity analysis* (paired or multiple spatial objects) to assess post-project geomorphic changes may require long observation periods (5–10 years) to capture high-flow years in which geomorphological changes are more likely to occur at a measurable level.

Basin-scale modelling (Benda and Dunne 1997; Coulthard *et al.* 2000) can use historical data to simulate and predict channel adjustments in response to basin changes. Geographical information system (GIS) databases can be combined with numerical modelling to reproduce sediment routing and its resultant changes in channel features (Montgomery *et al.* 1998) (see Chapter 15) and allowing better predictions of potential channel response, in terms of duration and extent and to characterize better longitudinal discontinuities and downstream changes in bed elevation, channel geometry, grain size and habitats.

From fluvial system to riverscape

Over the last decade, with the development of GIS technology and the increased availability of GIS information at the network scale, the spatial framework has been significantly enlarged and led to the concept of riverscape, considering the channel network as a complex set of geomorphic features connected and nested. The river style framework of Brierley and Fryirs (2005) and riverine ecosystem synthesis (RES) of Thorp *et al.* (2006) opened up this new research perspective in a conceptual way, whereas Alber and Piégay (2011) and Carbonneau *et al.* (2012) developed new technical procedures to characterize it quantitatively. Figure 5.17 shows the GIS procedure proposed to determine geomorphic features (e.g. channel reaches) characterized by specific properties (e.g. channel width significantly different between neighbouring reaches). This allows reconsideration of the cascading sediment system in a wider framework with new concepts for characterizing connectivity, such as buffers, barriers and blankets (Fryirs *et al.* 2007), and also with new technical issues to explore future channel changes as shown by the different scenarios of channel evolution and associated expected aquatic habitats developed by Bertrand *et al.* (2013) on the Drôme River. It relates geomorphic questions on forms and processes, notably cascading sediment transfers, with ecological

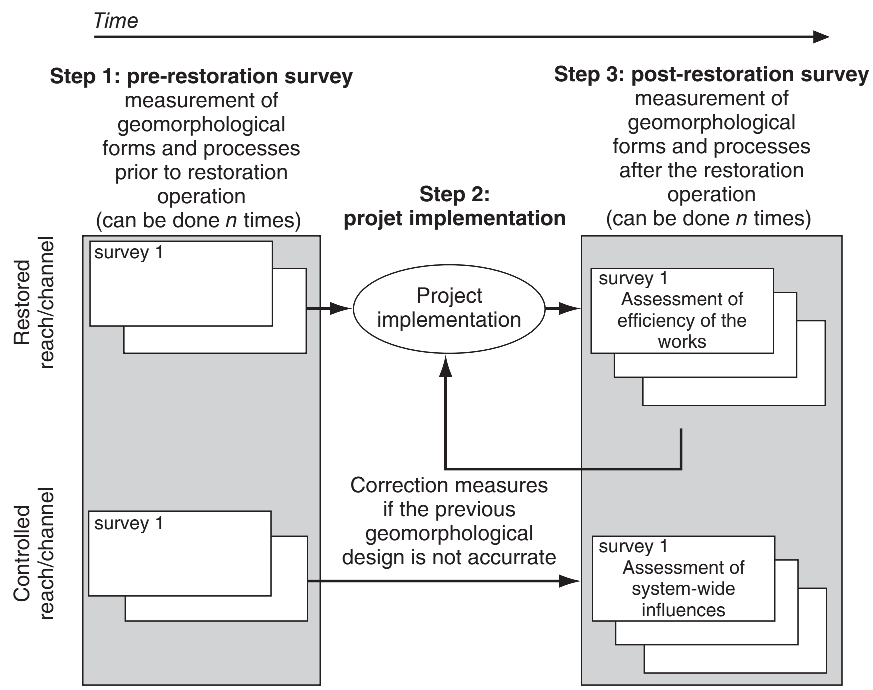


Figure 5.16 Application of similarity analysis in restoration projects. The restored site evolution is compared *n* times, at least one time before the implementation and one time after, with those of a control site unaffected in order to evaluate the efficiency of the measures done, removing the possible effects of factors affecting all the system and proposing corrected measures if the previous project does not reach the objectives. Source: Henry and Amoros, 1995. Reproduced with permission of Elsevier.

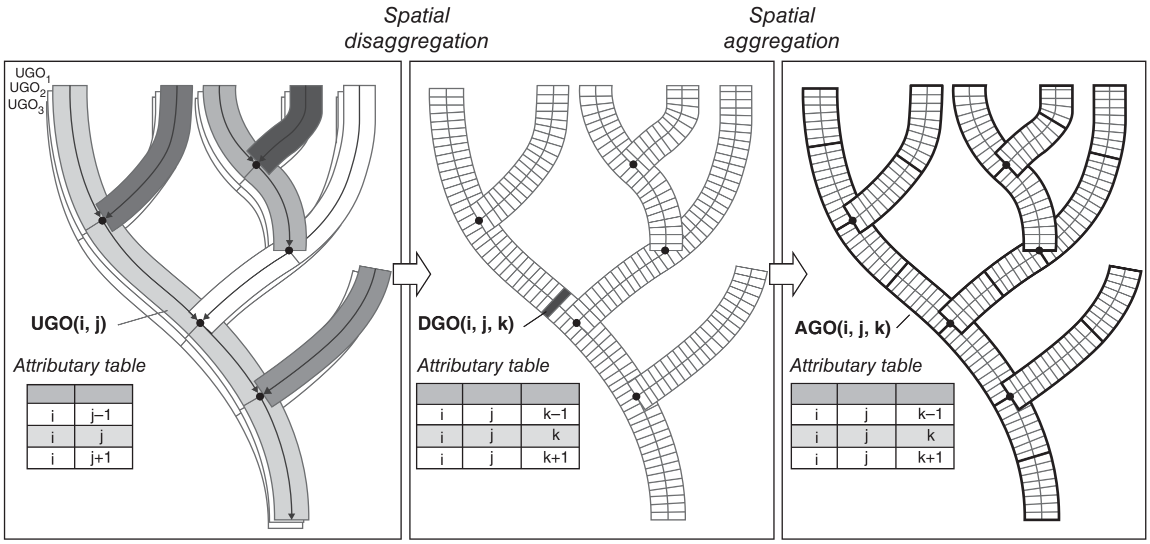


Figure 5.17 Conceptual framework of GIS procedure implemented at a network scale to determine geomorphic features at an appropriate spatial scale from spatial disaggregation and aggregation and showing the generic spatial units and the linear referencing axis for characterizing geographical objects UGO, DGO and AGO are respectively Unitary, Desaggregated and Aggregated Geographic Objects. Source: Alber and Piégay, 2011. Reproduced with permission of Elsevier.

issues linked to habitat assessment and associated ecological models (e.g. species presence or abundance for given physical conditions). Such approaches are emerging in ecology for characterizing riverine habitats at the network scale and targeting actions (see, for example, the use of a graph-based approach for targeting conservation effort by Eros *et al.* 2011).

5.4 Examples of applications

Research carried out based on the fluvial system concept commonly combines multiple approaches such as *similarity* (mainly LTS) and *connectivity* analysis. The most advanced approaches compare multiple nested components over time, as done for the Bega and Hunter Rivers in Australia (Brierley and Fryirs 2000, 2005) and in the French Pre-Alps (Liébault *et al.* 1999; Pont *et al.* 2009; Bertrand *et al.* 2013).

Bega River, Australia

Brierley and Fryirs (2000) and Fryirs and Brierley (2000) illustrated the importance of considering geomorphological adjustment to human impacts at a broad scale to frame river management and biological restoration. They used a total-system perspective to assess consequences of European settlement on fluvial forms of the Bega river (1040 km^2) on the south coast of New South Wales. The pre-settlement river was characterized by extensive swamps along the middle and upper reaches and a continuous low-capacity channel along the lowest gradient reaches. European settlement strongly modified hydrological regime, sediment supply and transfer and bank resistance, producing widespread channel widening and incision. To identify the character, capacity and stages of river recovery, comparative analysis (LTS) was based on retrospective analysis (e.g. archival plans, explorers' accounts, old ground and aerial photographs, hydrological data analysis to derive critical discharges) and on current field observations and measurements (long profiles and channel cross-sections, description of valley floor sedimentary structures, valley and channel morphology).

Channel features varied in space because of the internal characters of reaches (mainly valley morphology and distance downstream from the sediment sources) and in time because they have not reached the same adjustment stage at time *t*. Several homogeneous structural reaches were identified, within each of which the LTS model produced a distinct set of evolutionary stages: (i) the cut-and-fill river style, in wide, fully shaped valleys with steep slopes, (ii) the transfer style valley occupying the mid-basin reaches, bedrock confined with a lower gradient and a valley width up to 200 m and (iii) the floodplain accumulation river style in downstream reaches with a wide and low slope valley (Fig. 5.18a). With a detailed and fully documented evolutionary framework of river change and an appreciation of geomorphic linkages with a basin and associated limiting factors that may inhibit recovery potential, five stages of the LTS model were distinguished: the intact stage, the self-restored stage, the turning-point stage, the degraded stage and the created stage for which the character and the behaviour of the river reach do not equate to those of the predisturbance conditions) (Fig. 5.18b) (Fryirs and Brierley 2000). None of the narrow channels documented historically still exist. By 1900, the degradation and widening process was well advanced (real case B) and is still acting in the 1940s (real case C). The turning point occurred in the 1960s with island formations and exotic vegetation establishment (real case D). The authors expect

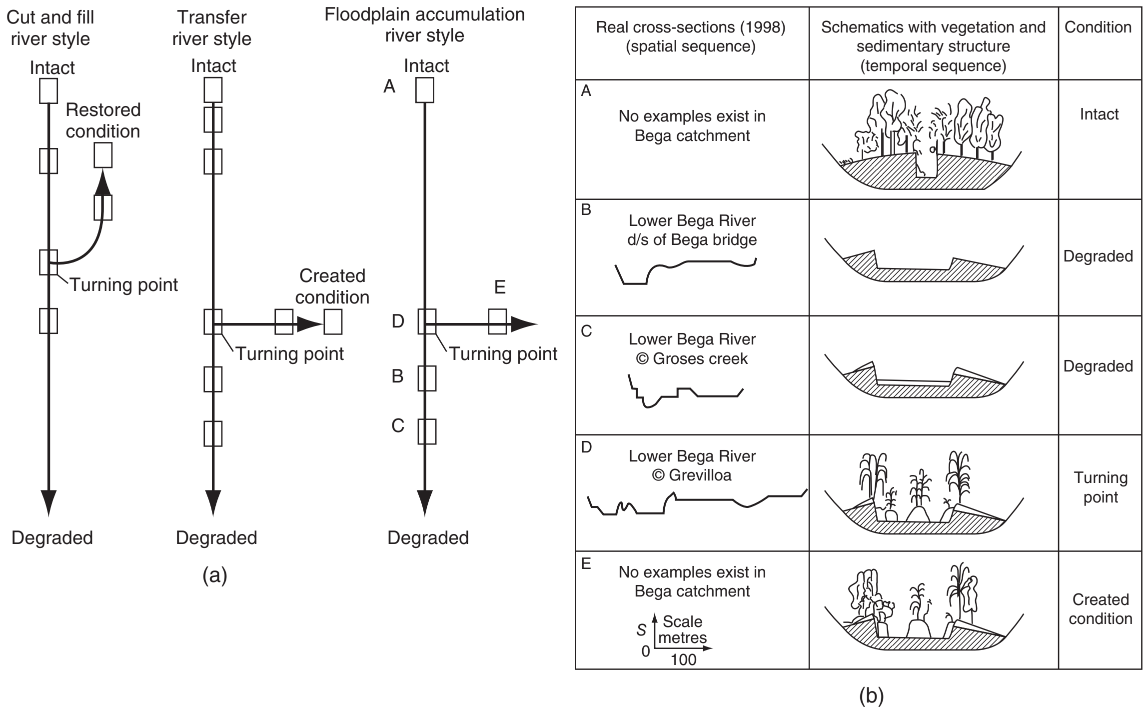


Figure 5.18 Conceptual model illustrating the temporal positions of different reaches of the Bega river system according to their style and the characters of the adjustment following human disturbances. Part (a) shows the different potential evolutionary stages according to the style and (b) gives stage examples concerning the floodplain accumulation river style. Modified from Fryirs and Brierley (2000).

created conditions (predicted case E) with a low-flow channel deepened and an increase in floodplain–channel connectivity with sediment removing along the channel bed (Fryirs and Brierley 2000).

The Drôme, Roubion and Eygues Rivers

Research on channel incision on the Drôme (1640 km^2), the Roubion (635 km^2) and the Eygues (1100 km^2) also illustrates the application of different conceptual tools presented in this chapter to understand the system evolution and to inform management decisions. Since 1994, an integrated analysis has been conducted on these systems located in the southern French Pre-Alps, focusing on multiple temporal and spatial scales and a wide range of tools (Fig. 5.19). The rivers all drain westward from limestone mountains under 2000 m in elevation, and flow into the middle Rhône River.

The approach utilized (Fig. 5.19) can be considered as an integrated or total-system analysis, as the geomorphological question is posed within basins, but with nesting information from sub-units. In this broad context, at each given spatial scale, comparisons are made between (i) the Drôme, the Eygues and the Roubion and (ii) the set of sub-basins, the tributary reaches, the main river segments (Fig. 5.19). The approach is then based both on *similarity analysis*, as single-scale components are compared (e.g. downstream alluvial reaches of the tributaries), and *connectivity analysis*, as the changes occurring on the upper levels are compared with those occurring at a lower level (downstream reach of tributaries in relation to their respective basins).

On the Drôme River, incision averaging 3 m was observed in downstream reaches where gravel mining was concentrated. Channel degradation has caused serious environmental problems, such as reduced channel dynamics and riparian vegetation regeneration, groundwater drawdown and undermining of levees and other infrastructure. River managers have recognized the need to assess the causes of the degradation beyond the reach scale and to manage bedload on longer time-scales than previously (e.g. decades instead of years) and over a larger spatial scale (i.e. from upper reaches to the Rhône instead of reach scale).

Enlarging the scope of analysis to the tributaries showed that the bedload supply from the basin was decreasing as a result of afforestation and erosion control since the 19th century (Landon *et al.* 1998). It was then necessary to understand the sediment transfer changes affecting the basin, to develop precise chronologies and analyse causes and to identify the active and potential sediment sources.

A *similarity analysis* was conducted on 50 sub-basins with field measurements (geometry and grain size analysis, scour chains and tracers in order to assess the bedload transport, erosion pins to evaluate the inner bedload input, ^{137}Cs and ^{210}Pb profiles but also dendrochronology to precise chronology of channel narrowing and deepening) in a historical perspective, aerial photographs taken in 1945, 1970 and 1995 and a land-survey

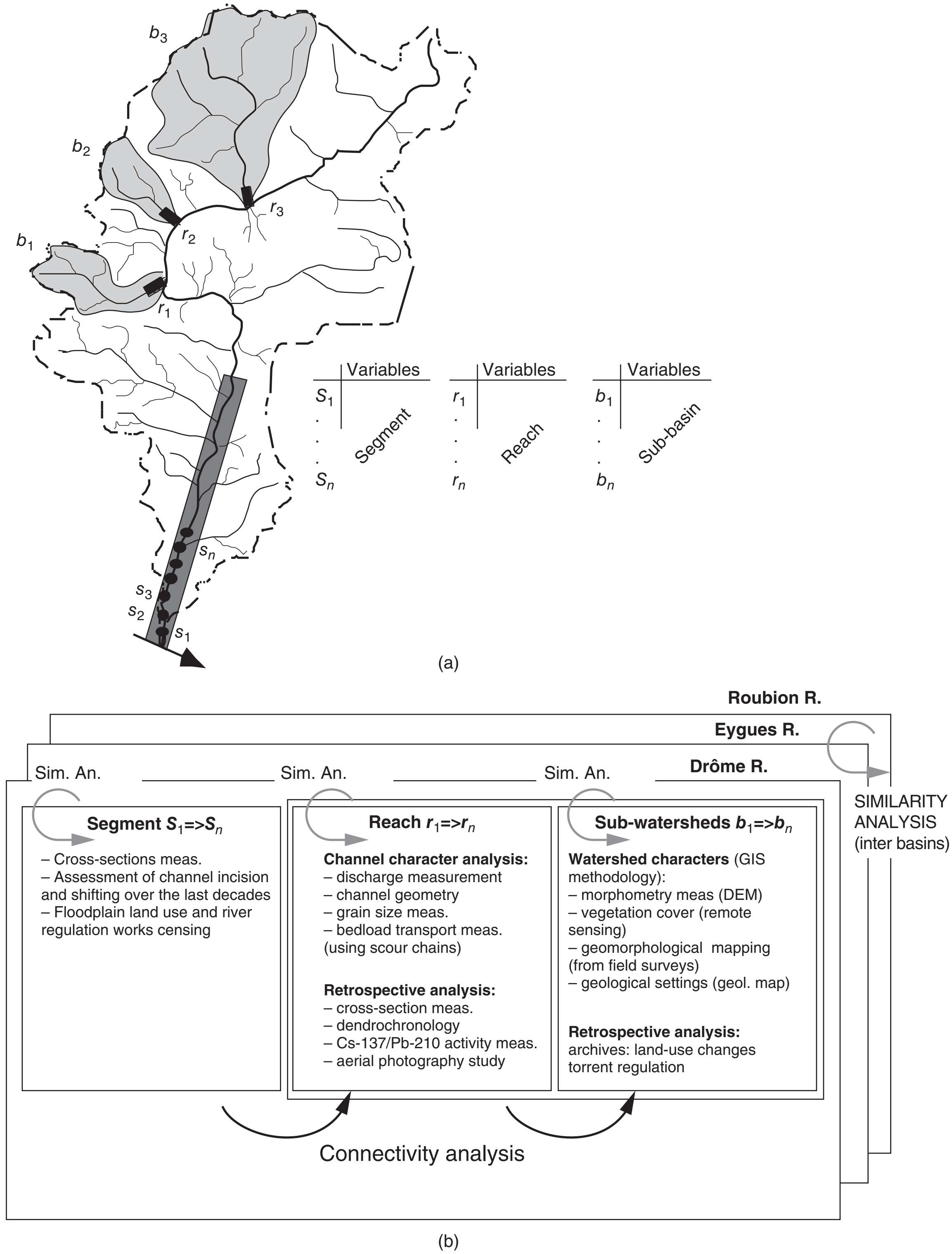


Figure 5.19 General schedule of geomorphological researches based on both *similarity* and *connectivity* analyses within the south Pre-Alps of France for assessing factors controlling the main stem changes. (a) Theoretical basin with nested components: tributary basins, tributary main reaches, main stem segments. (b) Data collection.

map of the middle 19th century being studied. A GIS developed from a digital elevation model (DEM), remote sensing images, aerial photographs and also archival data (maps, diagrams, written documents) was used to evaluate changes in vegetation and stream regulation and multivariate statistical analysis was performed to identify similarities among sub-systems (Fig. 5.20). Three fundamental phases of the approach are outlined in Table 5.1. Phase 1 is a detailed investigation of the study sites in the zone of sediment transfer. Phase 2 is an expansion of the study to adjacent landforms and to zone of sediment delivery. Phase 3 involves collection of historical information and the integration of the results of all three phases.

The *connectivity* analysis showed that tributaries still actively yielding high sediment loads are typically high gradient, with well-developed steep headwaters and many contacts between the stream network and highly erodible geomorphological units (see Fig. 5.20 for the Drôme case) (Liébault *et al.* 1999). The results indicated that a self-restoration process following the mining period was possible on the Eygues, but not on the Drôme. Even with a history of gravel mining comparable to the Drôme, the Eygues still experienced high bedload delivery from the basin, because the basin is more influenced by a Mediterranean climate (the vegetation cover is less extensive and the rainfall is more intense), and because of its geological setting, which resulted in a rapid transfer of sediment from the valley slopes to the channel. Of the three, the Roubion River has experienced the greatest reduction in sediment supply. A LCE analysis then indicated potential threshold factors explaining differences in channel adjustment among the three river systems.

5.5 Conclusions

A system approach is useful in fluvial geomorphological research, as it can provide a holistic framework within which to organize research, to understand sediment routing and to integrate sediment sources and their spatial and temporal variability (Table 5.2). The approach is also useful in river management as it permits hydraulic, hydrological, socioeconomic and ecological questions to be posed simultaneously and answered to solve interdisciplinary and applied problems.

The basin is recognized as the obvious unit for analysis and planning, a well-defined territory. Management at the basin level is increasingly recognized in the literature as necessary and increasingly adopted by government agencies, as illustrated by the European Water Framework Directive (adopted 2000), which requires hydrogeomorphic diagnosis and planning restoration measures performed at the river basin scale (so-called 'hydrographic districts'). The US Environmental Protection Agency has supported 'basin'- based planning, but so far these efforts are only encouraged and not universally effective.

In this new context of river management, a geomorphological approach allows better assessment of how and at what rate natural or human changes in a given part of a basin are likely to influence sedimentary and morphological features upstream and downstream (Newson 1994) and provide guidelines for restoration (Sear 1994; Kondolf and Downs 1996). Several authors have underlined the need to consider a geomorphological framework for biological improvement strategies (Sear 1994; Downs 1995; Brierley *et al.* 1999) (see also Chapter 21 for examples and detailed references) as well as engineering guidelines. Following Gilvear (1999), there are key contributions that fluvial geomorphology can make to the engineering profession with regard to river and floodplain management, such as promoting recognition of connectivity and interrelationships between river planform, profiles and cross-sections, stressing the importance of understanding fluvial history and chronology over a range of time-scales, highlighting the sensitivity of geomorphic systems to environmental disturbances and changes, especially when close to geomorphic thresholds. Physical habitats are often mapped in detail, but the temporal evolution of habitat mosaics, various spatial scales and the connectivity between components (nested perspective) must also be addressed in an interdisciplinary perspective to solve practical problems (Newson and Newson 2000).

Fluvial system approaches not only have advantages, but can also yield questionable conclusions when conducted without sufficient care or without adequate background (Table 5.2). When a good scientific practice is used, it can be very time consuming to collect sufficiently large data sets to describe different components of the system, and careful selection of samples is needed to develop an inferential statistical approach and support robust conclusions.

Acknowledgements

The research in the Rhône Basin system used to exemplify the chapter was carried out by a team that included G. Bornette, A. Citterio, F. Liébault and N. Landon and also the senior researchers of the former PIREN team of Lyon (particularly C. Amoros, J.P. Bravard, M. Coulet, G. Pautou and A.L. Roux), who developed the hydrosystem concept that provides the interdisciplinary framework for this work. The author also gratefully thanks G.M. Kondolf, D. Gilvear and G.E. Petts who reviewed the first edition of this chapter and brought useful comments to improve it, and also the late S.A. Schumm, who was a pioneer in this domain and contributed to the first edition of this chapter. Stan passed away in April 2011 and the author is greatly appreciative of his modesty, open-mindedness and kindness, allowing him as young scientist to share intense and fruitful discussions with him in 1998.

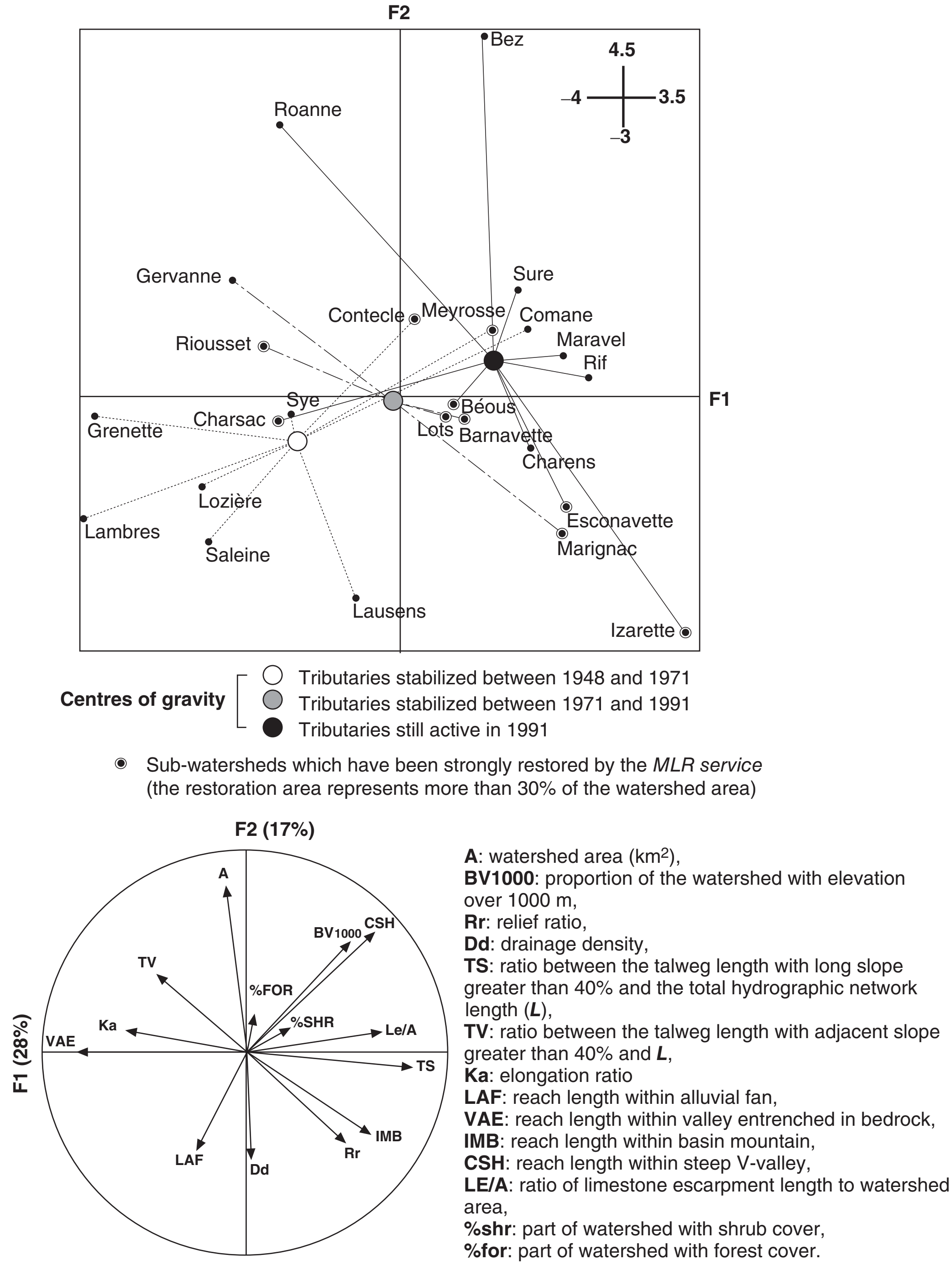


Figure 5.20 Example of integrated analysis performed on the Drôme basin to highlight basin characters which control tributary narrowing. The two graphs give the results of a normed Principal Component Analysis performed on 24 sub-basins with 14 morphometric, geomorphic and biogeographic variables. On the upper graph are projected the positions of the 24 tributaries grouped in three sets according to the chronology of their changes. On the lower graph, 'the correlation circle', are projected the variables measured on each tributary. Their directions allow interpretation of the position of the tributaries and the sets of tributaries in the graph above. From Liébault *et al.* (1999). Reproduced with permission of Arctic, Antarctic, and Alpine Research.

Table 5.1 List of elements to be considered in a coupled *similarity–connectivity* analysis, the example of the Drôme, Roubion and Eygues Basins.

Basic axis	Description
1. Present forms and processes within the study site	Geomorphic description of the reach (channel geometry and grain size), analysis of processes, e.g. assessment of reach conveying and trapping capacity (measurement of velocity, discharge and MES concentration, bedload transport, sedimentation rate within the floodplain, channel shifting and bank stability)
2. Spatial enlargement considering the floodplain/valley bottom and/or the basin	Study of floodplain (sedimentology and geometry, vegetation cover, land use) and basin characteristics (hydrographic network, basin morphometry, rainfall distribution, geology and vegetation patterns, sediment sources) from fieldwork and laboratory procedures (remote sensing and GIS analysis)
3. Temporal enlargement considering the channel, the floodplain/valley bottom and/or the basin	Study of changes over time in channel form and the variables listed above. At this stage, research may consider biological, physicochemical, geoarchaeological and sedimentary indicators but also archives (stream gauging records, plans for regulation of channel reaches, etc.). Written archives can be useful to understand the character and the chronology of land-use changes and some of the previous states of the system. The historical analysis should be conducted at a holistic scale, encompassing the nature and timing of changes affecting the neighbouring floodplain, the upstream channel network, or the whole basin

Table 5.2 Advantages and limitations of fluvial system approaches.

Main advantages

- Enlarge time and spatial scales when considering channel reach sensitivity allowing consideration of medium- and long-term changes and then consequences of human actions on river processes and forms
- Provide a conceptual framework for formulating hypotheses on channel evolution controls or critical processes, which can be tested by geomorphic tools such as experiments, mathematical modelling or GIS analysis
- Underline geographical complexity to understand limitations of reductionist approaches
- Formulate interdisciplinary questions and apply geomorphological knowledge for ecology and engineering purposes

Main limitations

- Based on empirical laws or expertise judgement, not necessarily on physical laws controlling river system, so that human experience and data available may significantly influence interpretations
- Errors in interpretation are common and risks of confounding facts and interpretation of facts are high
- Risk also of generalizing conclusions from a case-study to regional settings or from short-term (one-shot) field observation to a general understanding of driving factors
- Time consuming because it must cover a large area and uses multiple methods and materials (field data, documents, archives) to obtain robust conclusions

References

Agnelli, A., Billi, P., Canuti, P. and Rinaldi, M. 1992. Dinamica Morfologica Recente dell'Alveo del Fiume Arno, *CNR-GNDCI*, vol. 1739, Pisa: Pacini, 191pp.

Alber, A. and Piégay, H. 2011. Spatial disaggregation and aggregation procedures for characterizing fluvial features at the network-scale: application to the Rhône basin (France). *Geomorphology* 125(3): 343–360.

Amoros, C. and Bravard, J.P. 1985. l'Intégration du temps dans les recherches méthodologiques appliquées à la gestion écologique des vallées fluviales: l'exemple des écosystèmes aquatiques abandonnés par les fleuves. *Revue Française des Sciences de l'Eau* 4: 349–364.

Amoros, C. and Petts, G.E. 1993. *Hydrosystèmes Fluviaux*, Paris: Masson, 300pp.

Amoros, C., Richardot-Coulet, M. and Pautou, G. 1982. Les 'ensembles fonctionnels': des entités écologiques qui traduisent l'évolution de l'hydrosystème en intégrant la géomorphologie et l'anthropisation (exemple du Haut-Rhône français). *Revue de Géographie de Lyon* 57(1): 49–62.

Amoros, C., Bravard, J.P., Reygrobellet, J.L., Pautou, G. and Roux, A.L. 1988. Les concepts d'hydrosystème et de secteur fonctionnel dans l'analyse des systèmes fluviaux à l'échelle des écocomplexes. *Bulletin d'Écologie* 19(4): 531–546.

Benda, L.E. and Dunne, T. 1997. Stochastic forcing of sediment routing and storage in channel networks. *Water Resources Research* 33: 2865–2880.

Bennett, R.J. and Chorley, R.J. 1978. *Environmental Systems: Philosophy, Analysis and Control*, London: Methuen, 624pp.

Bertrand, M., Piégay, H., Liébault, F., Pont, D. and Sauquet, E. 2013. Sensitivity analysis of environmental changes associated with riverscape evolutions following sediment reintroduction: geomatic approach on the Drôme River network, France. *International Journal of River Basin Management* 11(1): 19–32.

Bravard, J.P. 1987. *Le Rhône, du Léman à Lyon*, Lyon: La Manufacture, 450pp.

Bravard, J.P., Amoros, C., Pautou, G., Bornette, G., Bournaud, M., Creuzé des Chatelliers, M., Gibert, J., Peiry, J.L., Perrin, J.F. and Tachet, H. 1997. River incision in south-east France: morphological phenomena and ecological effects. *Regulated Rivers – Research and Management* 13: 1–16.

Bravard, J.P., Kondolf, G.M. and Piégay, H. 1999. Environmental and societal effects of channel incision and remedial strategies. In: Darby, S.E. and Simon, A., eds., *Incised River Channels*, New York: John Wiley and Sons, pp. 303–341.

Bridge, J.S. 1993. The interaction between channel geometry, water flow, sediment transport and deposition in braided rivers. In: Best, J.L. and Bristow, C.S., eds, *Braided Rivers, Special Publication No. 75*, London: Geological Society of London, pp. 13–63.

Brierley, G.J. and Fryirs, K. 2000. River styles, a geomorphic approach to catchment characterization: implications for river rehabilitation in Bega catchment, New South Wales, Australia. *Environmental Management* 25(6): 661–679.

Brierley, G. J. and Fryirs, K.A. 2005. *Geomorphology and River Management: Applications of the River Styles Framework*, Oxford: Blackwell.

Brierley, G.J., Cohen, T., Fryirs, K. and Brooks, A. 1999. Post-European changes to the fluvial geomorphology of Bega catchement, Australia: implications for river ecology. *Freshwater Biology* 41: 839–848.

Brierley, G. J., Fryirs, K.A. and Hillman, M. 2008. River futures.In Brierley, G. J. and Fryirs, K.A., eds., *River Futures – An Integrative Scientific Approach to River Repair*, Washington, DC: Island Press, pp. 273–285.

Brooks, A. P., Brierley, G.J. and Millar, R.G. 2003. A paired catchment study between a pristine and a disturbed lowland alluvial river in southeastern Australia: channel morphodynamics before and after riparian vegetation clearance and wood removal. *Geomorphology* 51: 7–29.

Carbonneau, P., Fonstad, M.A., Marcus, W.A. and Dugdale, S.J. 2012. Making riverscapes real. *Geomorphology* 137(1): 74–86.

Carson, M.A. 1975. Threshold and characteristic angles of straight slopes. In: *Mass Wasting: Proceedings of the 4th Guelph Symposium on Geomorphology*, Norwich: GeoBooks, pp. 19–34.

Chin, A. and Gregory, K.J. 2001. Urbanization and adjustment of ephemeral stream channels. *Annals of the Association of American Geographers* 91(4): 595–608.

Chorley, R.J. and Kennedy, B.A. 1971. *Physical Geography: a Systems Approach*, London: Prentice-Hall.

Church, M. and Mark, D.M. 1980. On size and scale in geomorphology. *Progress in Physical Geography* 4: 342–390.

Citterio, A. and Piégay, H. 2009. Overbank sedimentation rates in former channel lakes: characterization and control factors, *Sedimentology* 56(2): 461–482.

Coulthard, T.J., Kirkby, M.J. and Macklin, M.G. 2000. Modelling geomorphic response to environmental change in an upland catchment. *Hydrological Processes* 14: 2031–2045.

Downs, P.W. 1995. River channel classification for channel management purposes. In: Gurnell, A.M. and Petts, G.E., eds, *Changing River Channels*, Chichester: John Wiley & Sons, pp. 347–365.

Dufour, S. and Piégay, H. 2009. From the myth of a lost paradise to targeted river restoration: forget natural references and focus on human benefits. *River Research and Applications* 25: 568–581.

Eros, T., Schmera, D. and Schick, R.S. 2011. Network thinking in riverscape conservation – a graph-based approach. *Biological Conservation* 144: 184–192.

Ferguson, R.I. 1986. Hydraulics and hydraulic geometry. *Progress in Physical Geography* 10: 1–31.

Friberg, N., Kronvang, B., Svendsen, L.M., Hansen, H.O. and Nielsen, M.B. 1994. Restoration of channelized reach of the river Gelsa, Denmark: effects on the macroinvertebrate community. *Journal of Aquatic Conservation: Marine and Freshwater Ecosystems* 4: 289–296.

Frissell, C.A., Liss, W.J., Warren, C.E. and Hurley, M.D. 1986. A hierarchical framework for stream habitat classification: viewing streams in a watershed context. *Environmental Management* 10: 199–214.

Fryirs, K. and Brierley, G.J. 2000. A geomorphic approach to the identification of river recovery potential. *Physical Geography* 21(3): 244–277.

Fryirs, K.A., Brierley, G.J., Preston, N.J. and Kasai, M. 2007. Buffers, barriers and blankets: the (dis)connectivity of catchment-scale sediment cascades, *Catena* 70(1): 49–67.

Gilvear, D.J. 1999. Fluvial geomorphology and river engineering: future roles utilizing a fluvial hydrosystems framework. *Geomorphology* 31: 229–245.

Glock, W.S. 1931. The development of drainage systems: a synoptic view. *Geographical Review* 21: 475–482.

Gregory, K.J. 1987. River channels. In: Gregory, K.J. and Walling, D.E., eds., *Human Activity and Environmental Processes*, Chichester: John Wiley & Sons, pp. 207–235.

Gregory, K.J., Davis, R.J. and Downs, P.W. 1992. Identification of river channel change due to urbanization. *Applied Geography* 12: 299–318.

Hack J.T. 1960. Interpretation of erosional topography in human temperate regions. *American Journal of Science* 258: 80–97.

Henry C.P. and Amoros C. 1995. Restoration ecology of riverine wetlands. *I. A scientific base. Environmental Management* 19(6): 891–902.

Hey, R.D. 1978. Determinate hydraulic geometry of river channels. *Journal of the Hydraulics Division, American Society of Civil Engineers* 104(6): 869–885.

Hey, R.D. and Thorne, C.R. 1986. Stable channel with mobile gravel-bed rivers. *Journal of Hydraulic Engineering* 8: 671–689.

Hooke, J.M. 1995. River channel adjustement to meander cutoffs on the river Bollin and river Dane, northwest England. *Geomorphology* 14: 235–253.

Huggett, R. 2007. A history of the systems approach in geomorphology. *Géomorphologie: Relief, Processus, Environnement* 2: 145–158.

Jacobson, R.B. and Berkley, J. 2011.Conceptualizing and communicating ecological river restoration. In: Simon, A., Bennett, S.J. and Castro, J.M., eds., *Stream Restoration in Dynamic Fluvial Systems, Geophysical Monograph Series 194*, Washington, DC: American Geophysical Union, pp. 9–27.

Junk, W.J., Bayley, P.B. and Sparks, R.E. 1989. The flood pulse concept in river-floodplain systems. *Canadian Special Publications of Fisheries and Aquatic Sciences* 106: 110–127.

Kirby, C. and White, W.R. 1994. *Integrated River Basin Development*, Chichester: John Wiley & Sons, 537pp.

Kondolf, G.M. and Downs, P.W. 1996. Catchment approach to planning channel restoration. In: Brookes, A. and Shields, F.D. Jr,, eds., *River Channel Restoration: Guiding Principles for Sustainable Projects*, Chichester: John Wiley & Sons, pp. 129–148.

Kondolf, G.M. and Piégay, H. 2011. Geomorphology and society. In: Gregory, K.J. and Goudie, A.S., eds., *The SAGE Handbook of Geomorphology*, London: SAGE Publications, pp. 105–117.

Kondolf, G.M., Piégay, H. and Landon, N. 2002. Channel response to increased and decreased bedload supply from land-use change contrasts between two catchments. *Geomorphology* 45: 35–51.

Landon, N., Piégay, H. and Bravard, J.P. 1998. The Drôme River incision (France): from assessment to management. *Landscape and Urban Planning* 43: 119–131.

Liébault, F. and Piégay, H. 2002. Causes of 20th century channel narrowing in mountain and piedmont rivers and streams of southeastern France. *Earth Surface Processes and Landforms* 27: 425–444.

Liébault, F., Clément, P., Piégay, H. and Landon, N. 1999. Assessment of bedload supply potentiality from the tributary watersheds of a degraded river: the Drôme (France). *Arctic, Antarctic and Alpine Research* 31(1): 108–111.

Meybeck, M. 2003. Global analysis of river systems: from Earth system controls to Anthropocene syndromes. *Philosophical Transactions of the Royal Society B: Biological Sciences* 358: 1935–1955.

Montgomery, D.R., Dietrich, W.E. and Sullivan, K. 1998. The role of GIS in watershed analysis. In: Lane, S.N., Richards, K.S. and Chandler, J.H., eds, *Landform, Monitoring, Modelling and Analysis*, Chichester: John Wiley & Sons, pp. 241–261.

Newson, M.D. 1994. Sustainable integrated development and the basin sediment system: guidance from fluvial geomorphology. In: Kirby, C. and White, W.R., eds., *Integrated River Basin Development*, Chichester: John Wiley & Sons, pp. 1–10.

Newson, M.D. and Newson, C.L. 2000. Geomorphology, ecology, and river channel habitat:mesoscale approaches to basin-scale challenges. *Progress in Physical Geography* 24(2): 195–218.

Orme, A.R. and Bailey, R.G. 1970. Effect of vegetation conversion and flood discharge on stream channel geometry: the case of Southern California watersheds. *Proceedings of the American Association of Geographers* 2: 101–106.

Paine, A.D.M. 1985. Ergodic reasoning in geomorphology: time for a review of the term? *Progress in Physical Geography* 9: 1–15.

Park, C.C. 1981. Man, river systems and environmental impacts. *Progress in Physical Geography* 5: 1–31.

Patton, P.C. and Schumm, S.A. 1975. Gully erosion, northern Colorado: a threshold phenomenon. *Geology* 3: 88–90.

Phillips J.D. 1996. Deterministic complexity, explanation, and predictability in geomorphic systems. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 315–335.

Piégay, H., Salvador, P.G. and Astrade, L. 2000. Réflexions relatives à la variabilité spatiale de la mosaïque fluviale à l'échelle d'un tronçon. *Zeitschrift für Geomorphologie* 44(3): 317–342.

Piégay, H., Pautou, G. and Bravard, J.P. 2003. L'histoire contemporaine des marges fluviales: entre renaturation et dénaturation. In: Pautou, G., Piégay, H., Ruffinoni, C., eds., *Les Ripisylves dans les Hydrosystèmes Fluviaux*, Paris: Institut pour le Développement Forestier, pp. 72–92.

Pont, D., Piégay, H., Farinetti, A., Allain, S., Landon, N., Liébault, F., Dumont, B. and Richard-Mazet, A. 2009. Conceptual framework and interdisciplinary approach for the sustainable management of gravel-bed rivers: the case of the Drôme River basin (SE France). *Aquatic Sciences – Research Across Boundaries* 71(3): 356–370.

Richards, K. 1996. Samples and cases: generalisation and explanation in geomorphology. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 171–190.

Schumm, S.A. 1973. Geomorphic thresholds and complex response of drainage systems. In: Morisawa, M., ed., *Fluvial Geomorphology*, Binghamton, NY: State University of New York, pp. 299–310.

Schumm, S.A. 1977. *The Fluvial System*, Chichester: John Wiley & Sons, 338pp.

Schumm, S.A. 1991. *To Interpret the Earth: Ten Ways to Be Wrong*, Cambridge: Cambridge University Press, 133pp.

Schumm, S.A. 1994. Erroneous perceptions of fluvial hazards. *Geomorphology* 10: 129–138.

Schumm, S.A., Kahn, H.R., Winkley, B.R. and Robbins, L.G. 1972. Variability of river patterns. *Nature Physical Science* 237: 75–76.

Schumm, S.A., Harvey, M.D. and Watson, C.C. 1984. *Incised Channels: Initiation, Evolution, Dynamics, and Control*, Littleton, CO: Water Resources Publications, 200 pp.

Schumm, S.A., Mosley, M.P. and Weaver, W.E. 1987. *Experimental Fluvial Geomorphology*, New York: John Wiley & Sons, 413pp.

Sear, D.A. 1994. Viewpoint: river restoration and geomorphology. *Aquatic Conservation: Marine and Freshwater Ecosystems* 4: 169–177.

Shields, F.D., Knight, S.S. and Cooper, C.M. 1997. Rehabilitation of warmwater stream ecosystems following channel incision. *Ecological Engineering* 8: 93–116.

Simon, A. 1989. A model of channel response in disturbed alluvial channels. *Earth Surface Processes and Landforms* 14: 11–26.

Thorp, J.H., Thoms, M.C. and Delong, M.D. 2006. The riverine ecosystem synthesis: biocomplexity in river networks across space and time. *River Research and Applications* 22(2): 123–147.

Townsend, C.R. 1989. The patch dynamics concept of stream community ecology. *Journal of the North American Benthological Society* 8: 36–50.

Trimble, S.W. 1997. Stream channel erosion and change resulting from riparian forests. *Geology* 25(5): 467–469.

Van Beek, J.L. 1981. Planning for integrated management of the Atchafalaya River basin: natural system viability and policy constraints. In: North, R.M., Dworsky, L.B. and Allee, D.J., eds., *Unified River Basin Management*, Minneapolis, MN: American Water Resources Association, pp. 328–337.

Ward, J.V. and Stanford, J.A. 1983. The serial discontinuity concept of lotic ecosystems. In: Fontaine, T.D. and Bartell, S.M., eds., *Dynamics of Lotic Ecosystems*, Ann Arbor, MI: Ann Arbor Science, pp. 29–42.

Wells, S.G., Bullard, T.F., Miller, J. and Gardner, T.W. 1983. Applications of geomorphology to uranium tailings silting and groundwater management. In: Wells, S.G., Love, D.W. and Gardner, T.W., eds., *Chaco Canyon Country: a Field Guide to the Geomorphology, Quaternary Geology, Paleoecology, and Environmental Geology of Northwestern New Mexico, American Geomorphological Field Group Field Trip Guidebook*, Albuquerque, NM: American Geomorphological Field Group pp. 51–56.

Whitlow, J.R. and Gregory, K.J. 1989. Changes in urban stream channels in Zimbabwe. *Regulated Rivers – Research and Management* 4: 27–42.

CHAPTER 6

Analysis of remotely sensed data for fluvial geomorphology and river science

David Gilvear[1] and Robert Bryant[2]
[1]University of Plymouth, Plymouth, UK
[2]University of Sheffield, Sheffield, UK

6.1 Introduction

Aerial photography and other remotely sensed data have increasingly been used as tools by the geomorphologist and river scientist. Remote sensing is based upon principles surrounding the transfer of energy from a surface to a sensor. Prior to the 1970s, the sensor, in the context of geomorphological mapping, was usually black and white photographic film and the platform an aeroplane. Since the early 1970s, however, there has been a huge increase in the number and spatial and spectral resolution of sensors and platforms (Tables 6.1 and 6.2) and accessibility of remotely sensed data offering the geomorphologist enhanced capabilities for interrogating the earth's surface. Remote sensing compared with traditional cartographic and field-based data collection has several advantages, including better spatial and temporal resolution, storage of data in digital format and interrogation of electromagnetic radiation (EMR), emitted or reflected, from land and water that is not detected by the human eye. A number of reviews have thus advocated the potential of using remote sensing as a tool to aid the investigation of rivers (e.g. Muller *et al.* 1993; Malthus *et al.* 1995; Milton *et al.* 1995; Lane 2000; Wealands *et al.* 2008; Marcus and Fonstad, 2008; Carbonneau and Piégay 2012). Given the impending launch of higher specification satellite sensors together with improvements in airborne sensors and digital camera and camcorder technologies, the future appears exciting in terms of gaining geomorphic coverage of rivers at multiple scales. The smallest of streams may also be interrogated at the reach scale using remotely sensed data acquisition methods via hand-held, tripod, crane or 'blimp' mounted sensors.

This chapter aims to provide a general review of the analysis of aerial photography and other remotely sensed data as a tool for studying fluvial processes and landforms with an emphasis on channel and floodplain environments. In particular, we aim to focus on remote sensing data taken from above-ground, aerial and space-borne remote systems. Techniques such as echo sounding, use of electrical resistivity and surface-based ground-penetrating radar are also forms of remote sensing but are not within the scope of this chapter. It is worth noting that remote sensing does not seek to replace traditional field-based methods of investigation, but rather to complement them by providing greater spatial coverage and in some cases greater temporal resolution, in each case giving access to a larger and more- ynamic sample population. Indeed, the real potential of applying remotely sensed data to fluvial research may only be realized if field-based methods are used to support remotely sensed data. For example, morphological data obtained at a cross-section on the ground can be extended to the reach and thence to the channel segment and finally the catchment scale. Overall, therefore, image analysis applied to remotely sensed data can potentially be used to provide information on hydrology, fluvial processes and spatial and temporal variability in land use at the catchment scale, thus putting riverine data into a landscape context. Indeed Whited *et al.* (2013) estimated juvenile salmon habitat from remotely sensed data on rivers within Alaska, British Columbia and the Kamchatka peninsula, encompassing an area of over 3 million km^2. Moreover, in the case of very large rivers (e.g. Amazon or Brahmaputra), viewing and capturing an image from the air is the only way to observe and quantify the overall morphology of the river. Furthermore, seeing the problem from a different viewpoint (literally) can provide new insights and suggest new hypotheses, which can then be tested in the field (Milton *et al.* 1995).

6.2 The physical basis

Photogrammetry

The two key geometric properties of an aerial photograph are angle and scale. According to the angle at which an aerial photograph is taken, it is referred to as either vertical, high oblique or low oblique. The following discussion relates to

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

Table 6.1 A selection of (a) currently or recently available data for high- and medium-resolution polar-orbiting platforms/sensors and (b) forthcoming high- and medium-resolution polar-orbiting platforms/sensors that are likely to be most relevant to geomorphological applications (for more details, see http://www.itc.nl/research/products/sensordb/).

(A)

Sensor name and platform	Launch date	Archive	Spectral bandwidths	Spatial resolution	Temporal resolution
SPOT 123/4 (HRV/HRVIR)	S1 1986	1986–2003	Pan: 510–730 nm (S123)	10 m	2–26 days
	S2 1990	1990–2009	Pan: 610–680 nm (S4)	10 m	depending on overlap coverage
	S3 1993	1993–1996	1: 500–590 nm	20 m	
	S4 1998	In orbit	2: 610–680 nm	20 m	
			3: 790–890 nm	20 m	
SPOT 5 (HRG)	2002	In orbit	Pan: 480–710 nm	5 m (2.5 m)	2–3 days
			1: 500–590 nm	10 m	
			2: 610–680 nm	10 m	
			3: 790–890 nm	10 m	
			4: 1580–1750 nm	20 m	
Landsat 5 (MSS)	1986	1972–	1: 500–600 nm	80 m	16–18 days
Landsat 4	1982		2: 600–700 nm		
Landsat 3	1978		3: 700–800 nm		
Landsat 2	1975		4: 800–1100 nm		
Landsat 1	1972				
Landsat 6 (ETM)	1993 (failed)	1993–1993	1: 450–520 nm	30 m	16 days
Landsat 5 (TM)	1984	In orbit	2: 520–600 nm	30 m	
Landsat 4 (TM)	1982	1982–1993	3: 630–690 nm	30 m	
			4: 760–900 nm	30 m	
			5: 1550–1750 nm	30 m	
			6: 10400–12500 nm	120 m	
			7: 2100–2350 nm	30 m	
Landsat 7 (ETM+)	1999	In orbit (faulty scan line as of 2003)	Pan: 520–900 nm	15 m	16 days
			1: 450–515 nm	30 m	
			2: 525–605 nm	30 m	
			3: 630–690 nm	30 m	
			4: 750–900 nm	30 m	
			5: 1550–1750 nm	30 m	
			6: 10400–12500 nm	60 m	
			7: 2090–2350 nm	30 m	
EO-1 (Hyperion)	2000	2000–2011	Hyperspectral 400–2500 nm 220 bands with a 10 nm resolution	30 m	Variable
EO-1 (ALI)	2000	2000–2011	Pan: 480–690 nm	10 m	Variable
			MS-1*: 433–453 nm	30 m	
			MS-1: 450–510 nm	30 m	
			MS-2: 525–605 nm	30 m	
			MS-3: 630–690 nm	30 m	
			MS-4: 775–805 nm	30 m	
			MS-4*: 845–890 nm	30 m	
			MS-5: 1200 –1300 nm	30 m	
			MS-6: 1550 –1750 nm	30 m	
			MS-7: 2080–2350 nm	30 m	
ALOS (PRISM) (AVNIR)	2006	2006–2011	PRISM 520–770 nm	2.5 m	2–46 days
			AVNIR		
			Blue: 420–500 nm	10 m	
			Green: 520–600 nm	10 m	
			Red: 610–690 nm	10 m	
			Near-IR :760–890 nm	10 m	

Table 6.1 (*continued*)

(A)

Sensor name and platform	Launch date	Archive	Spectral bandwidths	Spatial resolution	Temporal resolution
DMC (Disaster Monitoring Constallation)	2002/3	In orbit	Pan: 520–900 nm	4 m	Potentially 1 day
	2005/10		Green: 520–600 nm	32 (22) m	
			Red: 630–690 nm	32 (22) m	
SlLIM-6 (SlLIM-6-22)			Near-IR: 770–900 nm	32 (22) m	
Ikonos	1999	In orbit	Pan: 450–900 nm	0.8 m	~3 days
			Blue: 450–530 nm	3.2 m	
			Green: 520–610 nm	3.2 m	
			Red: 640–720 nm	3.2 m	
			Near-IR: 770–880 nm	3.2 m	
GeoEye-1	2008	In orbit	Pan: 450–800 nm	0.41 m	<3 days
			Blue: 450–510 nm	1.65 m	
			Green: 510–580 nm	1.65 m	
			Red: 655–690 nm	1.65 m	
			Near-IR: 780–920 nm	1.65 m	
Orbview 3	2003	2003–2007	Pan: 450–900 nm	1 m	<3 days
			Blue: 450–520 nm	4 m	
			Green: 520–600 nm	4 m	
			Red: 625–695 nm	4 m	
			Near–IR: 760–900 nm	4 m	
Quickbird 2	2001	In orbit	Pan: 450–900 nm	0.6 m	2.5–5.6 days depending on latitude
			Blue: 450–520 nm	2.4 m	
			Green: 520–600 nm	2.4 m	
			Red: 630–690 nm	2.4 m	
			Near–IR: 760–890 nm	2.4 m	
Rapideye Constallation	2008–	In orbit	Blue: 440–510 nm	6.5 m	
			Green: 520–590 nm	6.5 m	
			Red: 630–685 nm	6.5 m	
			Red edge: 690–730 nm	6.5 m	
			Near-IR: 760–850 nm	6.5 m	
Cartosat 1 (IRS-P5)	2005	In orbit	Pan: 500–850 nm	2.5 m	4–5 days
Cartosat 2	2007	2007–2012	Pan: 500–850 nm	0.8 m	
Cartosat 2A	2008	In orbit	Pan: 500–850 nm	0.8 m	
Cartosat 2B	2010	In orbit	Pan: 500–850 nm	0.8 m	
IRS-1C/D (LISS3)	1995 (1C) 1997 (1D)	1988–	LISS3	5.8 m	5–24 days depending on latitude
			Pan: 500–800 nm	23.5 m	
			Green: 520–590 nm	23.5 m	
			Red: 620–680 nm	23.5 m	
			Near-IR: 770–860 nm	23.5 m	
			SWIR: 1550–1700 nm		
Resources at 1 and 2	2003/2011	2003–	LISS4	5.8 m	5 days
LISS3/4			520–590 nm	5.8 m	
			620–680 nm	5.8 m	
			770–860 nm		
WorldView 1	2007	In orbit	Pan: 450–800 nm	0.5 m	1.7–5.9 days

(*continued overleaf*)

Table 6.1 (*continued*)

(A)

Sensor name and platform	Launch date	Archive	Spectral bandwidths	Spatial resolution	Temporal resolution
WorldView 2	2009	In orbit	Pan: 450–800 nm	0.46 m	
			Coastal: 400–450 nm	1.85 m	
			Blue: 450–510 nm	1.85 m	
			Green: 510–580 nm	1.85 m	
			Yellow: 585–625 nm	1.85 m	
			Red: 630–690 nm	1.85 m	
			Red edge: 705–745 nm	1.85 m	
			Near-IR 1: 770–895 nm	1.85 m	
			Near-IR 2: 860–1040 nm	1.85 m	
FORMOSAT-2	2004	In orbit	Pan: 450–900 µm	2 m	1 day
			Blue: 450–520 nm	8 m	
			Green: 0.52–0.60 µm	8 m	
			Red: 0.63–0.69 µm	8 m	
			Near-IR: 0.76–0.90 µm	8 m	
EROS-A/B	2000 A	In orbit	Pan: 450–900 nm	1.90 m A	1–15days
	2006 B	In orbit		0.70 m B	
SPIN-2	1999	1999–	TK-350 camera	10 m	Variable
			Pan: 510–760 nm	2 m	
			KVR-1000 camera		
			Pan: 510–760 nm		
Terra and Aqua	2000 (Terra)	2000–	36 spectral bands ranging in wavelength from 0.4 to 14.4 µm	VIS 250 m	1 day
(EOS AM-1 and PM-1)	2002 (Aqua)	Terra		NIR 500 m	
MODIS		2002– Aqua		TIR 1 km	
MERIS (ENVISAT)	2001	In orbit	15 programmable bandwidths with a spectral resolution of 2.5 nm ranging from 390 to 1040 nm	300–1200 m	2–3 days

(B)

Sensor name and platform	Launch date	Spectral bandwidths	Spatial resolution	Temporal resolution
SPOT 6/7 (NAOMI)	2012–2013	Pan: 0.51–0.73 nm	2 m	1–5 days
		Multispectral	8 m	
Pléiades 1 and 2	2011–2012	Pan: 480–830 nm	0.5 m	1–3 days
		Blue: 430–550	2 m	
		Green: 490–610	2 m	
		Red: 600–720 nm	2 m	
		Near-IR: 750–830 nm	2 m	
GeoEye-2	2013	Pan: 450–800 nm	0.25 m	<3 days
		Blue: 450–510 nm	<1.65 m	
		Green: 510–580 nm	<1.65 m	
		Red: 655–690 nm	<1.65 m	
		Near-IR: 780–920 nm	<1.65 m	
Landsat 8	2012	Pan: 500–680 nm	14 m	16 days
Landsat Data		Coastal: 433–453 nm	28–30 m	
Continuity Mission		Blue: 450–515 nm	28–30 m	
(LDCM)		Green: 525–600 nm	28–30 m	
		Red: 630–680 nm	28–30 m	
		Near-IR: 845–885	28–30 m	
		Cirrus: 1360–1390 nm	28–30 m	
		SWIR 1: 1560–1660 nm	28–30 m	
		SWIR 2: 2100–2300 nm	28–30 m	

Table 6.1 (*continued*)

(B)

Sensor name and platform	Launch date	Spectral bandwidths	Spatial resolution	Temporal resolution
NPOESS Preparatory Project (NPP) (VIIRS)	2011	36 spectral bands ranging in wavelength from 402 to 11800 nm	370–740 m	1 day
Worldview 3	2014	Pan: 450–900 nm	0.31 m	4–5 days
		Blue: 450–510 nm	1.24 m	
		Green: 510–580 nm	1.24 m	
		Red: 630–690 nm	1.24 m	
		Near-IR: 770–895 nm	1.24 m	
		Coastal: 400–450 nm	1.24 m	
		Yellow: 585–625 nm	1.24 m	
		Red edge: 705–745 nm	1.24 m	
		Near-IR 2: 860–1040 nm	1.24 m	
Cartosat 3	2011	Pan: 500–750 nm	0.3 m	<5 days
		Blue: 450–520 nm	<1 m	
		Green: 520–590 nm	<1 m	
		Red: 620–680 nm	<1 m	
		Near-IR: 770–860 nm	<1 m	
Formosat 5	2013	Pan: 450–900 μm	2 m	1–3 days
		Blue: 450–520 nm	4 m	
		Green: 0.52–0.60 μm	4 m	
		Red: 0.63–0.69 μm	4 m	
		Near-IR: 0.76–0.90 μm	4 m	
EnMAP	2013	Hyperspectral: 420–1000 nm (94 bands) 900–2500 nm (134 bands)	30 m	Unknown
PRISMA	2013	Pan:		Unknown
		400–700 nm	2.5–5 m	
		Hyperspectral:		
		400–2500 nm (10 nm bands)	20–30 m	

Table 6.2 Examples of common operational airborne multispectral and hyperspectral remote sensing systems. See http://www.eufar.net/ and http://arsf.nerc.ac.uk/ for a detailed list of airborne sensors and platforms.

Sensor name	Acronym	Spectral coverage (nm)	No. of available wavebands
Airborne Visible/Infrared Imaging Spectrometer	AVIRIS	410–2450	224
Reflective Optics System Imaging Spectrometer	ROSIS	430–880	28
Multispectral Infrared and Visible Imaging Spectrometer	MIVIS	400–2500	92
Modular Airborne Imaging Spectrometer	MAIS	440–2500	71
		7800–11800	7
CCD Airborne Experimental Scanner for Applications in Remote Sensing	CAESAR	520–780	9
Digital Airborne Imaging Spectrometer	DAIS	400–2500	72
ITRES – Compact Airborne Spectrographic Imager	CASI	410–925	288 or 15
As above	CASI 1500	950–2450	100
As above	CASI 550	8–11.5 μm	32
ITRES Hyperspectral SWIR Imaging System	SASI 600	3700–4800	
ITRES Pushbroom Hyperspectral Thermal Sensor System	TASI 600		
ITRES Thermal Airborne Broadband Imager	TABI 1800		
Daedalus/Argon 1268 ATM	ATM	420–2350	10
		850–13000	1
Daedalus/Argon AHS - Airborne Hyperspectral Scanner	AHS	443–13384	80
Specim AISA Eagle	AisaEAGLE	400–970	200
Specim AISA Hawk Hyperspectral Instrument	AisaHAWK	970–2450	185
Specim AISA DUAL	AisaDUAL	400–2500	320

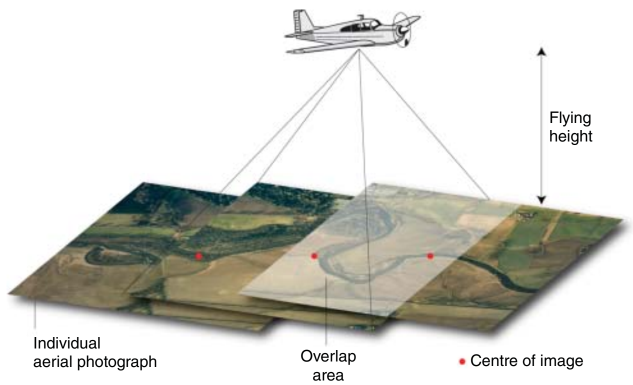


Figure 6.1 The principles of aerial photograph acquisition from an aircraft to allow subsequent photogrammetry and quantification of channel and floodplain topography.

vertical aerial photography. Vertical aerial photographs are normally taken in sequences along an aircraft's flight line with an overlap of usually more than 60% to allow the photographs to be viewed three-dimensionally or stereoscopically (Fig. 6.1). A small-scale aerial photograph will provide a synoptic, low spatial resolution overview (e.g. 1 : 50,000) of a large area; such a photograph may be useful for mapping drainage networks but is only appropriate for detailed reach-scale analysis of river morphology on large rivers. A large-scale aerial photograph will provide a high spatial resolution view of a small area; such a photograph will be useful for detailed analysis of a reach but if data for a long length of river are needed, it would entail the use of a large number of such aerial photographs. The scale of a photograph is determined by the focal length of the camera and the vertical height of the lens above the ground.

Overlapping pairs of aerial photographs can provide a three-dimensional view of the Earth's surface by the effect of parallax. Parallax refers to viewing an object from two different angles; humans use the principle by focusing on an object with their left and right eyes. In the case of aerial photographs, optical devices called stereoscopes are used to view a pair of stereo aerial photographs and the ground appears to the viewer to be in three dimensions. The phenomenon of parallax can be used to measure the height of objects. Parallax results in points of higher elevation having a greater horizontal displacement on successive aerial photographs than a lower elevation feature. The value of parallax displacement is positively related to the distance between the centre of the two photographs and the height of the object of interest and negatively related to the height above the ground from which the photograph was taken (Fig. 6.1). Modern computer-based photogrammetry allows automated production of digital terrain models from stereo aerial photography and such techniques are obviously important for the subject of geomorphology (Lane *et al.* 1993). More detail on the potential of analytical and digital photogrammetry in geomorphological research can be found in Lane *et al.* (1993), Dixon *et al.* (1998) and Chandler (1999). An excellent review of photogrammetric applications for the study of channel morphology and associated data quality issues can also be found in Lane (2000).

Electromagnetic radiation and remote sensing systems

EMR reflected from the Earth's surface can vary with location, time, geometry of observation and waveband (Verstraete and Pinty 1992) (Fig. 6.2). Consequently, the successful interpretation of remotely sensed data for a particular river will depend upon an understanding or characterization of these four factors. In particular, an understanding of the way in which EMR interacts with the surface of the Earth and what factors affect its capture by a sensor is important. Many good reviews of the nature and interaction of EMR and Earth surface features exist (e.g. Asrar 1989). Aspects relevant to fluvial geomorphology are summarized here.

Electromagnetic radiation

EMR occurs as a continuum of wavelengths. The wavelengths of greatest interest when remotely sensing the Earth are the reflected radiation in the visible and near and middle infrared wavebands, emitted radiation in the middle and thermal infrared wavebands and reflected radiation in the microwave wavebands. EMR originates from a source; this is usually the Sun's reflected light or the Earth's emitted heat but can be man-made as in active microwave radar. Initially, EMR passing through the atmosphere may be distorted and scattered. In general, greater scattering and distortion occur with greater distance between the Earth and sensor and the greater the levels of atmospheric moisture, pollutants and dust. Generally, atmospheric noise is wavelength specific and can be easily removed by ignoring those wavelengths that are affected (e.g. for hazy image scenes caused by Rayleigh scattering, short wavelengths can be omitted from the image set). However, some atmospheric effects (e.g. Mie and non-selective scatter of EMR) are more difficult to remedy or take account of (Kaufman 1989; Cracknell and Heyes 1993). Overall, the level of correction undertaken for atmospheric effects can depend on whether qualitative or quantitative data are to be extracted from imagery. For the latter, correction and calibration using in situ (alternatively called ground-truthed) data are commonly necessary.

Once EMR interacts with the surface, one of three processes can occur: (i) reflection of energy, (ii) absorption of energy and (iii) transmission of energy. In general, the amount and characteristics of each of these three energy interactions will depend upon the inherent characteristics of the Earth's surface and the wavelength of EMR that is interacting with it. For example, visible wavelengths are reflected from water in a different way than those wavelengths in the near and middle infrared regions. Consequently, in order to generate geomorphologic information successfully from remote sensing data, a knowledge of how EMR interacts with the specific surfaces is needed. Most objects can only be differentiated if the reflectance from the surface is different from that of the adjacent object in the wavebands being

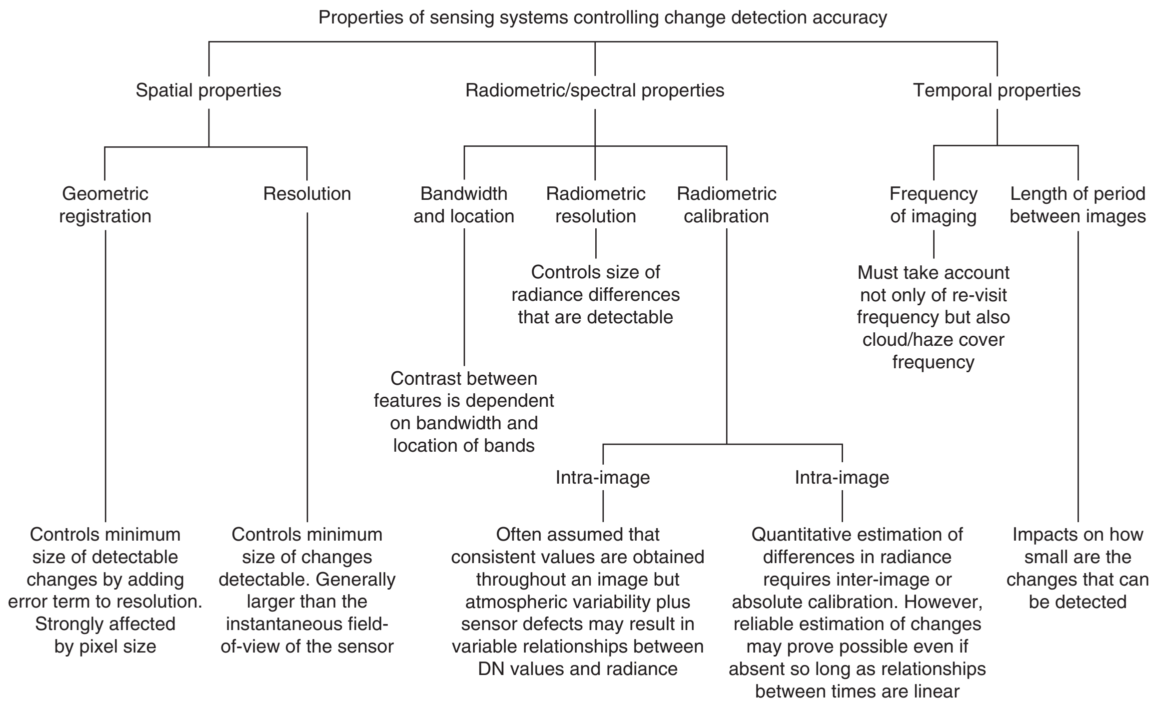


Figure 6.2 The main properties of remote sensing systems controlling the accuracy of temporal change and spatial variability detection. Source: Townshend and Justice, 1988. Reproduced with permission of Taylor and Francis.

captured by the sensor and above the radiometric precision of the sensor. In most cases, such information can be obtained from either the literature (e.g. Irons *et al.* 1989), spectral libraries (e.g. online libraries such as at: http://speclib.jpl.nasa.gov) or field collection of reflectance spectra coincident with image acquisition using a spectroradiometer. A brief review of the spectral properties of surfaces within the fluvial realm is outlined later.

Sensors and platforms

Most sensors commonly may have several channels capturing information in narrow, broad or continuous bandwidths. Generally, sensors with few channels (1–10) and broad bands (50–100 nm) are referred to as multispectral. Sensors with the capability of measuring in numerous (up to 250 bands), narrowly defined (to 1–10 nm) bands or continuous parts of the electromagnetic spectrum are referred to as hyperspectral. Different sensors may also have different radiometric resolutions, which will control the size of radiance differences at the Earth's surface that can be detected. Different sensors normally also capture different components of EMR (Tables 6.1 and 6.2).

Given a knowledge of the reflectance properties of fluvial surfaces, it is important to understand how these data will be recorded at the sensor (Tables 6.1 and 6.2). Most photographic sensors/cameras differ from hand-held cameras only in that they have dedicated film magazines, automated drive mechanisms and a large supporting lens cone. Similarly, the cameras can record data using common film types. For visible wavelengths, either black and white panchromatic film or true colour film can be used. For other wavelengths, black and white and false colour near-infrared film may also be utilized. Photographic sensors can produce hard-copy images using either (1) strip, (2) panoramic or (3) frame formats. Digital sensors can essentially have two different types of design, either (i) an optical mechanical scanner (or multispectral across-track scanner) or (ii) a linear array (i.e. an along-track push-broom of charge-coupled devices). In addition, sensors of each type have a predetermined spatial resolution (the edge length of a square or rectangular land parcel from which an individual signal can be deduced; see later for more detail) and swath width (visible area on each pass). Useful reviews of imaging spectrometry can be found in Curran (1994) and Plummer *et al.* (1995). It should also be noted that recent advances in combining the output of global positioning systems (GPSs) with image capture on a variety of platforms has increased the potential for accurate identification of absolute location on the Earth's surface. Similarly, now that geomorphic data are routinely geo-referenced with GPS (e.g. Milne and Sear 1997), these can be linked to individual pixels on imagery, permitting more accurate image calibration and validation.

Sensors can be further characterized by their platforms. These can range from a satellite to aircraft or even balloons. For most existing and forthcoming satellite platform–sensor combinations, the repeat period can range from 12 hours to 44 days (Table 6.1). In the case of airborne sensors, a greater temporal flexibility can be afforded (Table 6.2). The pros and cons of airborne data versus satellite data for river research are given in Table 6.3.

Table 6.3 The pros and cons of airborne data versus spaceborne data for fluvial geomorphology.

	Airborne photography*	Airborne imaging spectrometer†	Airborne multi-spectral scanner‡	Spaceborne multispectral scanner§
Resolution				
Spatial range (m)	<0.5	0.5–20	0.5–20	10–80
Spectral bandwidth (nm)	~10	1.8	5–20	60
Radiometric range (DN)	Hard copy	4092	256–4029	256
Temporal repetition (days)	Upon request	Upon request	Upon request	3–18
Logistics				
Number of spectral bands	1–3	288	1–15	7
Swath width	Small¶	Small¶	Small¶	Large
Mission targeting	Upon request	Upon request	Upon request	None or limited
Flight time	Upon request	Upon request	Upon request	Fixed
Repetitive coverage	Low cost	High cost	High cost	Default (lower cost)
Cost per km^2	Low	High	High	Lower
State-of-the-art technology	No	Most recent	Most recent	10–15 year lag
Atmospheric influence	Low	Less	Less	Highest
Sky conditions required	Less critical	Critical	Critical	Extremely critical
Sensor platform motions	Roll, pitch and yaw	Roll, pitch and yaw	Roll, pitch and yaw	Negligible
Hands-on repairs/adjustments	In situ	In situ	In situ	Almost impossible
Standardized products	All	Few	Some	More

*E.g. Wild RC10 Survey Camera (black-and-white panchromatic, black-and-white infrared, colour and colour infrared).
†E.g. CASI, AVIRIS.
‡E.g. Daedalus 1268 ATM.
§E.g. SPOT, Landsat TM.
¶Depends on the flying height.
Modified from Dekker *et al.* (1995).

Considerations

The most important considerations when acquiring imagery for a particular site are whether the radiometric resolution of the sensor, the amount of atmospheric scatter, the surface roughness of the objects and the spatial variability of reflectance within the wider field of view can affect the ability to differentiate between objects. The latter factor is important because the radiance recorded from an area of ground also contains radiance from surrounding areas. Another important factor to be aware of is that the raw digital number (DN) values often need to be calibrated to radiance units and this calibration may not be constant across an image or between images if atmospheric distortion or illumination is variable. Even after calibration, some wavebands may have to be discarded owing to high nois-to-signal ratios or simply to poor calibration (e.g. Bryant and Gilvear 1999). Uneven radiation capture at the sensor, due to variations in scene illumination, equally applies when scanning aerial photographs to allow image processing. Hence Gilvear *et al.* (1995) needed to apply shade correction to scanned aerial photographs of different parts of Faith Creek in Alaska to detect meso-scale habitat change. Shade correction was necessary because of differences in natural illumination (i.e. atmospheric conditions) at the time the photographs were taken, uneven illumination when the photographs were scanned (using a video camera system) and differences in photographic processing. The grey-tone in an aerial photograph when captured in digital format is assigned an 8-bit DN value between 0 (black) and 255 (white) according to its grey-tone. This number of grey-tones is much greater than the human eye can detect, allowing image analysis to identify spatial variability that would go undetected with manual observation.

Scale and spatial accuracy issues

Size of river

One of the main considerations in using remote sensing to study the fluvial geomorphology of rivers is channel width. A number of studies of large rivers (>200 m wide) have been undertaken using satellite remote sensing (e.g. Salo *et al.* 1986; Phillip *et al.* 1989; Ramasamy *et al.* 1991) and more recently during space shuttle missions (see later). Milton *et al.* (1995) suggest that for smaller rivers (~20–200 m wide), airborne remote sensing, incorporating high-resolution advanced sensors and improved temporal/spatial flexibility, may be a more suitable approach for mapping and monitoring change. For very small rivers (<20 m wide), a hand-held helium blimp or model aircraft with remotely operated camera or oblique imagery, that is subsequently rectified, may be more appropriate in gaining the spatial resolution of imagery required (Carbonneau *et al.* 2012).

Image format

One key characteristic of aerial photography and satellite imagery is that it provides typically a square format (whether digital or hard-copy). Unfortunately, this does not match well with rivers, which form linear features in the landscape. When these images are used, the length of river shown will not be much greater than the edge length of the image unless it has high sinuosity (Fig. 6.3a). Therefore, to cover an appreciable river length a number of images often have to be pieced together to form a mosaic unless very small-scale images (e.g. >1 : 25,000) are used. Even with large-scale photographs covering a small reach, with small- to medium-sized rivers channel features may be hard to detect given the small area that the river covers on the image, especially where riparian woodland obscures some of the channel. This mismatch in geometry results in increased costs in the purchase of imagery, increased time for image rectification and more ground control points.

A distinction should also be highlighted here between aerial photography flown specifically for the purpose of a riverine study with systematic coverage of a whole region or country. In the first case, overlapping aerial photographs will be oriented along the direction in which the river is flowing and this will maximize the length of river on each photograph. In the second case, the photographs will not be oriented parallel or be focused on the river and only small lengths of the river will be found on some of the appropriate photographs. Coverage of countries or regions is also typically small-scale and if large-scale, the number of flight lines needed to permit full coverage may become excessively large. Aerial photography that follows a watercourse (unless it is very wide and very small-scale photography is needed) requires only one flight line and larger scale photography is therefore more acceptable. In the US regional surveys of 1 : 20,000 or 1 : 25,000 scale for example, might be commissioned by the US Geological Survey or Department of Agriculture, whereas 1 : 12,000 scale or better photography might be commissioned by the US Army Corps of Engineers or Bureau of Reclamation (e.g. for flood hazard mapping), but may or may not give good coverage of riparian and floodplain areas. With digital data and appropriate image analysis software, one can zoom in on areas of the channel of interest to gain a large-scale picture, but this may not be appropriate if the spatial resolution of the image is too small. However, airborne remote sensing captures digital data with a fixed swath width, for a given flying elevation, but with infinite length. Thus data can be captured, for example, as a 1 km by 10 km area, covering perhaps a 15 km length of a medium-sized river together with its floodplain. It is therefore ideally suited to collecting data on rivers (Milton *et al.* 1995).

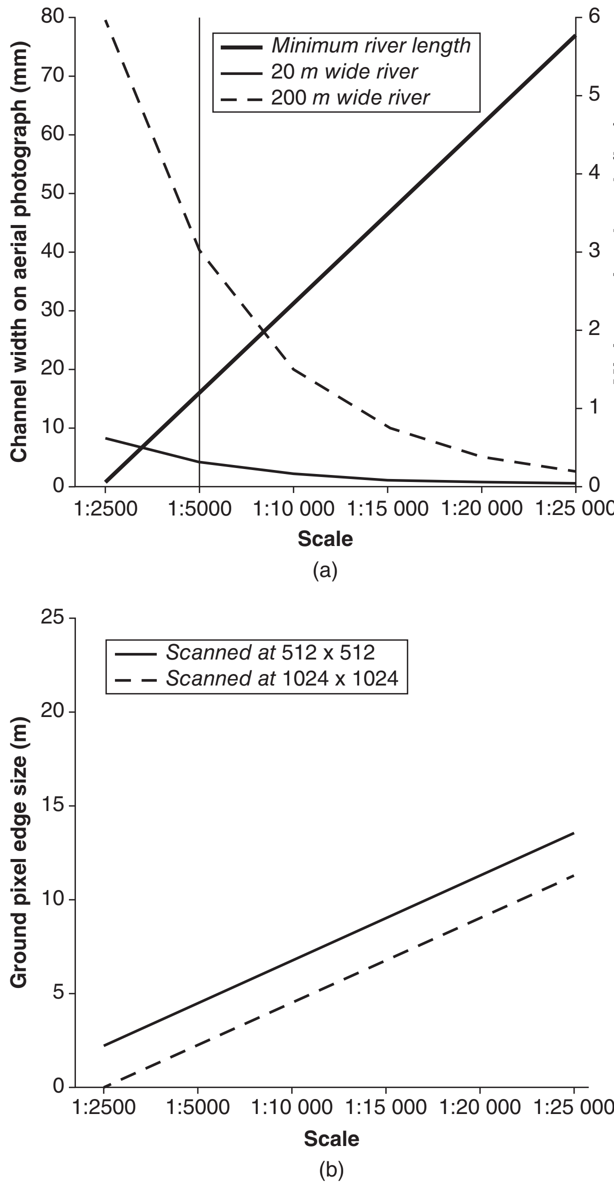


Figure 6.3 The effects of image scale, scanning resolution and river size on spatial resolution. (a) Aerial photograph scale, minimum river length observable and width of the channel on the photograph for 20 and 200 m widths. (b) Ground pixel resolution on scanned aerial photographs for differing scales and scan resolution.

Sensor resolution

The spatial resolution of a sensor is usually described by a distance in metres, which relates to the edge length of a single square or rectangular parcel of land from which a radiation value can be assigned (a pixel). Pixel size relates to sensor type and altitude. The higher the platform, the larger is the pixel size for a given sensor and the wider the swath width (Fig. 6.4), although this can vary depending on the sensor optics and the size of the charge-coupled devices (CCDs) used. In scanning aerial photographs and producing digital imagery, one must also calculate the pixel size in relation to the photographic scale and scanning resolution (Fig. 6.3b). This will limit the amount of detail and minimum size of object that can be detected. Although one can enlarge a particular area of a photographic print or zoom in on a digital image to gain greater detail, there

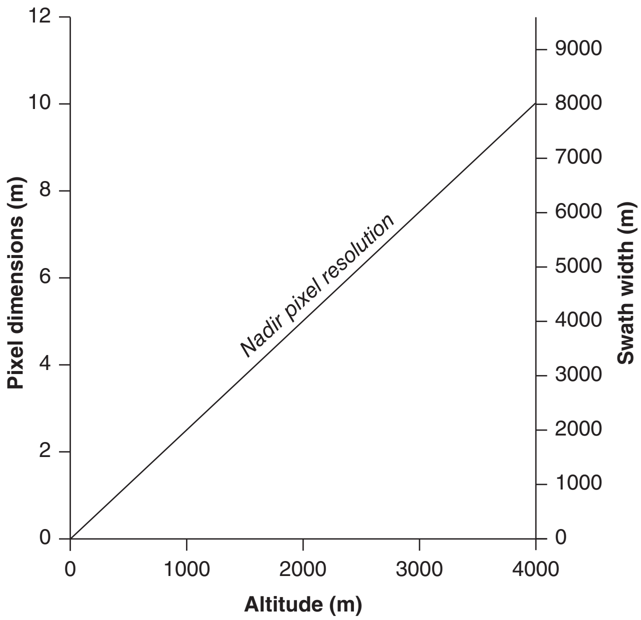


Figure 6.4 Relationship between pixel size and swath width with platform altitude for the Daedalus 1268 airborne thematic mapper (ATM) scanner. Modified from Wilson, 1995.

must come a point where no further information can be visually obtained, particularly on an image made up of pixels. The grain size of the photographic film may also, but only rarely, limit the minimum size of fluvial features that can be detected (Lane *et al.* 1994). Grain size is the theoretical minimum and scanning density the practical minimum.

A key question in remote sensing is whether the pixel size is smaller than the object (e.g. landform) of interest. If larger, identification of a landform will be difficult and delimitation of its boundaries impossible. Even if the pixel size is smaller than the landform of interest, problems can still arise because normally a number of pixels are needed to identify a feature effectively. First, pixel edges may not necessarily coincide with the edge of features on the ground. Therefore, many pixels will contain a number of objects (mixed pixels). In the case of the fluvial environment this could be bed material, water and vegetation. The combination of these reflectances may result in the pixel appearing similar to another type of surface and hence be misinterpreted. In the case of multispectral imagery, the problems of mixed pixels can be met using mixture modelling to estimate the proportion of each spectral 'end-member' present within a pixel (Mertes *et al.* 1993; Bryant 1996). Second, problems arise because a scanner not only receives radiation from the area of ground demarcated by pixel edges (the ground resolution element – GRE) but also surrounding areas (instantaneous working area – IWA). Indeed, even within a GRE, the scanner will not respond uniformly to radiation from its area because pixel intensities are not independent but auto-correlated. More detail on what a pixel means in reality can be found in Cracknell (1998). On a scanned black and white aerial photograph, individual pixels of soil and water can have the same DN value. A density sliced image would therefore assign them to a similar land cover type when from visual observation of the image the difference in surface type would be obvious because of the pixels' location in the larger image and contextual information. In this regard, feature integrity needs to be incorporated into the classification procedure used. Image analysis should therefore not always be seen to be superior to visual interpretation but rather as a complementary approach. Of course, with colour aerial photographs, soil, water and vegetation surfaces are more easily distinguished, but distinguishing pure water and areas with submergent and emergent aquatic plants may be difficult.

Geometric accuracy

Spatial resolution and scale control precision and should not be confused with accuracy in that the image may not be geometrically correct and may include tilt and warping. To rectify images, ground control points (GCPs), for which a *relative* or absolute location is known, have to be matched with the corresponding feature on the image using a mathematical transformation. In the consideration of temporal change and an absence of GCPs, image registration to each other using objects that are known not to have moved can be undertaken. In some remote areas, with no man-made objects, such identification can be problematic. If the chosen objects move (e.g. bank lines due to erosion), results can be spurious. Satellite platforms are highly stable (i.e. they remain perpendicular to the ground surface and do not suffer roll and pitch as with an aircraft) and often a first-order transformation based on relatively few GCPs is sufficient to gain a high level of accuracy. Airborne platforms are often less stable, but techniques to correct photographic images geometrically from nadir and oblique pointing cameras are well developed (Lane *et al.* 1993; Chandler 1999). Successful rectification of airborne scanner data may require the survey area to be split into smaller sections (e.g. Christensen *et al.* 1988) or the transformation to be localized (e.g. Devereux *et al.* 1990), or the use of a parametric correction procedure based on aircraft altitude measurements (e.g. Wilson 1997). It may benefit from the incorporation of a digital elevation model (Cosandier *et al.* 1994), but probably not in level, relatively flat floodplain environments. Information on spatial accuracy and error in relation to aerial photography is provided in Chapter 4. Townshend and Justice (1988) have reviewed the properties of remote sensing systems that control the accuracy of land cover assessments, focusing on spatial aspects (Fig. 6.2). They emphasize the importance of matching the spectral, radiometric and spatial capabilities of the sensor with the properties of the surfaces being sensed. Moreover, the timing and frequency of sensing must coincide with that of temporal change or events within the fluvial environment, and even then, in the case of many sensors, cloud cover may prevent observation.

Spectral properties and the fluvial environment

Landscape components

Figure 6.5 shows typical spectral response curves, measured using airborne thematic mapper (ATM) data, for surfaces and features found in the fluvial environment. It is apparent that spectral responses normally fall into three distinct classes (Hooper 1992; Milton *et al.* 1995): (i) water, shadow and aquatic vegetation; (ii) trees and other green vegetation; and (iii) exposed sediment. It should be noted, however, that in the case of black and white aerial photographs and some other spectral wavebands, these classes might not be so clearly differentiated. These three classes may be thought of as the 'spectral end-members' of the riverine environment. In the context of fluvial geomorphology, the two main end-members are water and sediment. Interrogation of subtle differences in the radiation from water and exposed sediment can reveal more about their nature and hence the relationship between these surfaces and electromagnetic radiation is explored further below. Vegetation is another component that is often of interest to the fluvial geomorphologist in that floodplain vegetation mosaics often relate to topography and soils and channel mobility (Hickin and Nanson, 1975).

As mentioned earlier, the majority of visible, near and middle infrared radiation reaching a soil or sediment surface is either reflected or absorbed and little is transmitted. The five characteristics of sediment, which are interrelated and which determine its reflectance properties, are, in order of importance: moisture content, organic content, texture, structure and iron oxide content. Most important is the relationship between soil moisture and reflectance. Reflectance decreases substantially in the visible wavelengths with increasing moisture content until soil reflectance becomes saturated. Reflectance in the near and middle infrared wavelengths is also negatively correlated with soil moisture and an increase in soil moisture will result in rapid falls in reflectance, particularly in wavelengths centred at 0.9, 1.4, 1.9, 2.2 and 2.7 nm. Moisture will have a greater effect on the reflectance of clay soils than sandy sediments. Soil organic matter will also decrease reflectance up to a content of 4–5%.

The complex relationship between fluvially deposited sediment and spectral characteristics is demonstrated by the results of Bryant *et al.* (1996) and Rainey *et al.* (2000) (Fig. 6.6). Figure 6.6 shows that immediately before a high tide there is a positive reflectance between ATM band 9 reflectance and increasingly particle size, but immediately after the high tide and thorough wetting of the inter-tidal sediments there is a negative relationship. Such knowledge of changes in spectral reflectance–physical substrate properties with differing degrees of wetness is obviously vital to the sound interpretation of remotely sensed data and also illustrates the need for concomitant field-based measurements. It also demonstrates that a sound knowledge of the spectral characteristics of the features of interest and how they respond to environmental variables can also help in optimizing the use of remotely sensed data.

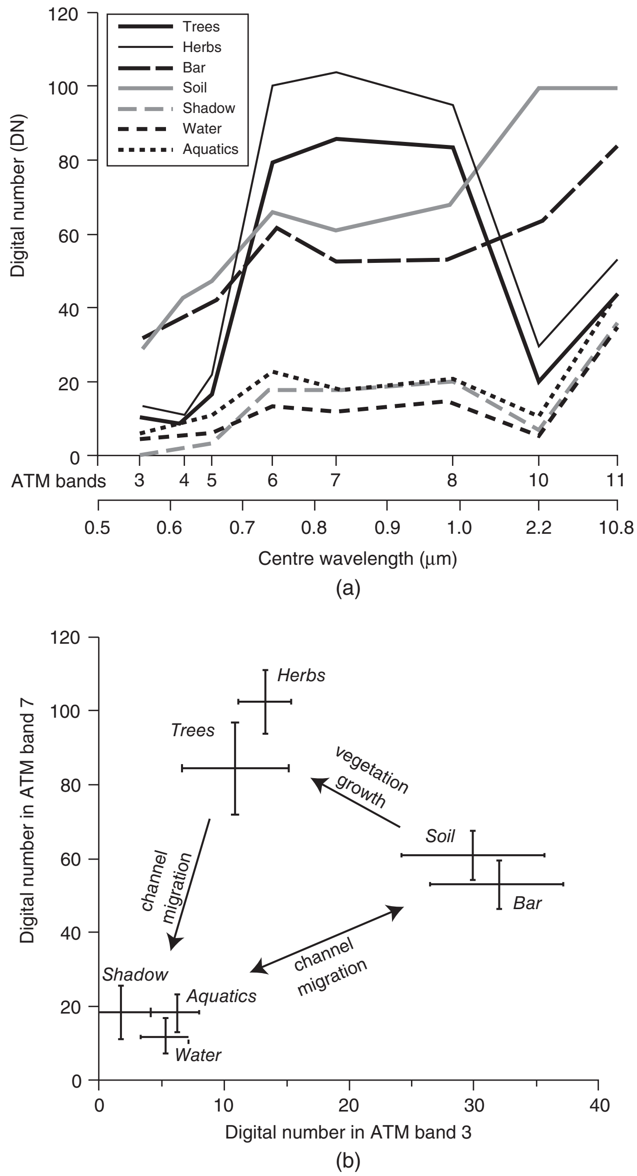


Figure 6.5 Spectral signatures for the fluvial environment based on imagery of the River Teme, UK. (a) Representative spectral response curves ascertained from ATM data. (b) Differentiation of fluvial environments using the range of ATM band 7 and band 3 values for each surface type and their relationship to channel change. Modified from Hooper, 1992 and Milton *et al.*, 1995.

The radiance of thermal infrared wavelengths from a soil is primarily determined by its moisture content. The wetter the soil is, the cooler it will be during the day and the warmer it will be at night. Soils and sediment in general generate a low radar return, and only when they are recorded at moderate to low incidence angles do they generate a moderate return and are sensitive to soil moisture variations.

Unlike soil and sediment, the majority of visible, near and middle infrared radiation is either absorbed or transmitted at pure water surfaces. In visible wavelengths, little light is

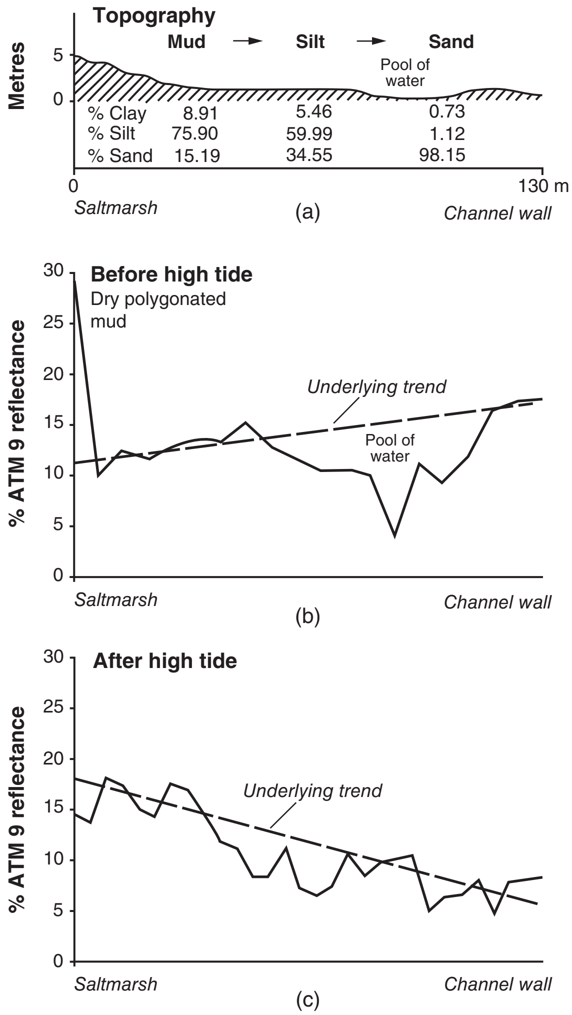


Figure 6.6 Variable ATM reflectance from inter-tidal sediments either side of a high tide on the River Ribble, England, showing that moisture content not only results in a shift in reflectance but also reverses the relationship between reflectance and particle size. (a) Inter-tidal morphology and sedimentology. (b) Spatial variation in spectral reflectance of the inter-tidal sediments before the high tide. (c) Spatial variation in spectral reflectance of the inter-tidal sediments after the high tide and re-wetting. Source: Rainey *et al*, 2000. Reproduced with permission of Taylor and Francis.

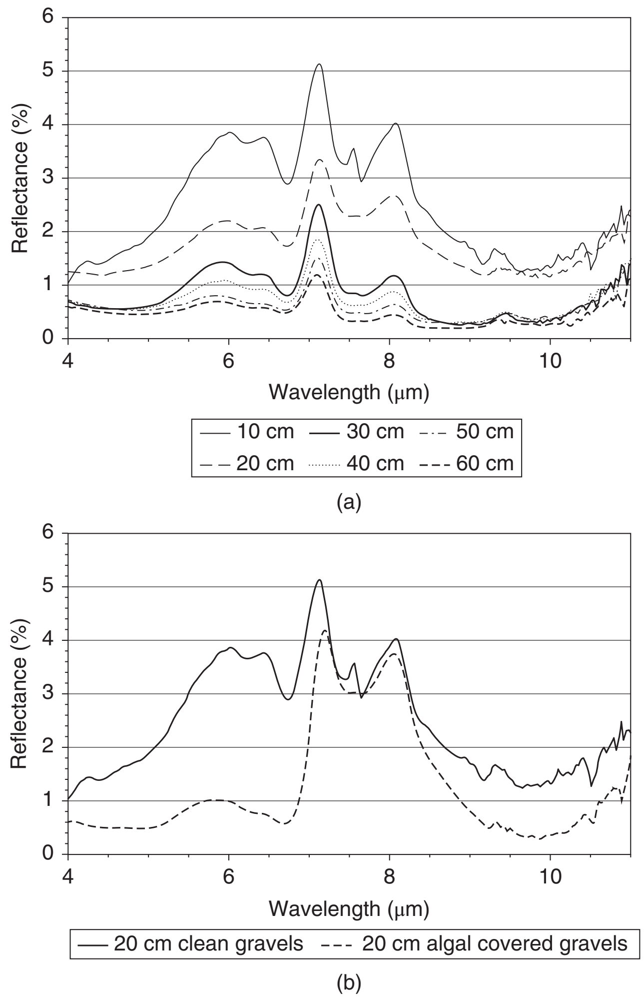


Figure 6.7 Spectral reflectance as recorded on the River Tummel, Scotland, using a field spectrometer showing the affect of (a) water depth and (b) bottom types. After Gilvear *et al*, 1997.

absorbed, a small amount (usually under 5%) is reflected and the majority is transmitted. Water also absorbs near and middle infrared wavelengths strongly. This results in sharp contrasts between water and land boundaries, with pure water appearing black, for instance, on infrared aerial photographs. The factors that affect the spatial variability in the reflectance are depth of water, the suspended and solute content of the water and the surface roughness of the water. In shallow water, some radiation is reflected not by the water itself but by the substrate. Therefore, in shallow water it is often the channel bed that determines the water's reflectance properties and colour, in the absence of high suspended sediment loads or colour levels.

The effects of differing water depths and differing substrates on spectral reflectance, as measured using a field spectrometer, are shown in Fig. 6.7. Water has a similar thermal inertia to soil and yet it has a much smaller diurnal thermal range. It may therefore be differentiated by being warmer at night and cooler during the day, with differences most marked early in the morning. Thermal imagery, especially where collected in the early hours of morning, is therefore often best used in differentiating soil, water boundaries and soil moisture variability. A water body is usually an area of low radar return and appears black on radar images, although speckling may occur if waves are present and at right-angles to the radar pulse.

In the case of the spectral reflectance from an image or part of an image of particular interest not encompassing the full range of DN values, contrast stretching can be applied to heighten differences in the image. Thus, given that the grey-tone variability across the water surface was limited on scanned aerial photographs of Faith Creek, Alaska, a contrast stretch was applied solely to the water surface (Gilvear *et al.* 1995; Winterbottom and Gilvear 1997) to allow a stronger link to water depth to be extrapolated. Other parts of the image were masked off and the lightest grey-tone (equivalent to the shallowest areas) was assigned a value of 255 and the darkest areas (equivalent to the deepest areas) a value of zero. This allowed the differences in water depth in the image to be more clearly identified. Many other image enhancement techniques are available to improve the 'quality' of images but are outside the scope of this chapter.

Classification of imagery into Earth surface categories based on their spectral properties can be undertaken using a variety of methods. For more detailed types of classification and further information on enhancement and classification techniques, a textbook on remote sensing should be consulted (e.g. Sabins 1996). Raw DN values or radiation fluxes can be converted to an environmental variable (e.g. water depth or soil moisture content) using an empirically based relationship derived with ground-truth data or an established algorithm. Supervised classification of pixels is based upon assigning pixels with similar spectral characteristics to pixels identified in training areas. Training areas are one or more pixels assigned a land surface category because the cover type is known from field survey. Unsupervised classification relies on methods that assign pixels to a predetermined number of groups according to spectral similarity. This may or may not relate to land cover types or features on the ground.

Summary overview

The preceding sections have demonstrated that there are many factors to be taken into account when selecting a remote sensing approach to interrogate a fluvial system. The key variables that will determine the choice of data and approach taken are: (i) the length of river being studied; (ii) the width of the river; (iii) the spatial resolution required; (iv) the land cover or water surface or sub-surface properties to be detected; (v) the degree of precision and accuracy acceptable; and (vi) the frequency with which changes might need to be detected. Following consideration of these variables, the fluvial geomorphologist may then be able to determine whether a ground-based, airborne or spaceborne approach will be most appropriate and whether single spectra or multispectral or hyperspectral data are required. Table 6.5 provides a generic protocol for assessing the most suitable remote sensing approach according to the type of fluvial study being undertaken

6.3 River geomorphology and in-channel processes

2D channel morphology and channel change

Two-dimensional mapping of river channel morphology and channel change has been the focus of fluvial geomorphology for a number of decades (Table 6.4). In recent years, the synoptic mapping of channel planforms has been revolutionized by the availability of Google Earth (Lisle 2006), which has created a unique new resource for 2D mapping of channel morphology and floodplains. The key issue is always scale, however, whatever the nature of the remotely sensed data. Large channels are easy to observe on aerial photographs and satellite data using a range of imagery at all scales. The only constraint is restricted river lengths on large-scale images and thus high purchase costs and time involved in producing maps of drainage networks for large areas. For example, the anastomosing channels on the Niger delta have been mapped with satellite data (Diakite *et al.* 1986; Brivo *et al.* 2002). As the channel becomes smaller, however, spatial resolution becomes critical. France *et al.* (1986), working in Wales, concluded that Landsat TM data could record lakes as small as 0.6 ha and streams down to 3–5 m in width with acceptable accuracy. Thirty-three first-order streams were thus detected. However, 1 : 10,000 aerial photograph interpretation revealed 156 first-order streams, many of which were less than 1.0 m wide. For the delineation of ephemeral streams in Nevada, 76% of second-order and larger streams could be identified in SPOT panchromatic images (Gardner *et al.* 1987). At a smaller scale and in a more complex situation, Schumann (1989) identified the 'parent' channel and the relative importance of anabranches in an anastomosing reach of Red Creek, Wyoming, using black and white aerial photographs. The parent channel was darkest due to grasses and sagebrush flanking the channel where moisture availability was highest.

The effect of spatial scale and re-sampling regime on planform detection from remotely sensed data is covered in Chapter 4. In the case of the spatial resolution of the imagery being used to its limit, a waveband that achieves the greatest contrast between land and water is most suitable (e.g. near and middle infrared). Increasingly robust automated classification of channels will become possible (Argialas *et al.* 1988). Many other examples of imagery being used to map channel planform could be cited, but this is not necessary given the straightforward and obvious simplicity of the technique. However, problems can sometimes occur in detection and bankfull definition, and here the expertise of the geomorphologist is of paramount importance.

In-channel features have also been mapped extensively using aerial photographs and satellite imagery. Aerial photography, for example, has been used to map bar forms for over five decades as part of channel planform studies (e.g. Werritty and Ferguson 1980; Warburton *et al.* 1993). Similarly on large rivers, satellite imagery has been used to map bar morphology (Thorne *et al.* 1993). Figure 6.8, for example, clearly shows overall channel planform and bar morphology of the Mississippi River above Vicksburg. The image, covering an area of about 28 km by 21 km was acquired in October 1994 by spaceborne imaging radar. Imagery such as that shown in Fig. 6.8 can be used easily to produce quantitative data on geomorphic attributes such as channel width and size and shape of exposed channel bars. However, the bar size and shape as depicted on the image is stage dependent and successive images cannot always be compared directly unless water levels are known to be the same at each of the

Table 6.4 Recent examples of the application of remote sensing to fluvial geomorphology and river science.

Location	**Geomorphic purpose**	**Imagery type, and scale/pixel size**	**Reference**
(a) 2D channel morphology and channel change			
Mekong River, Thailand–Lao PDR	Bank erosion	Aerial photography and Spot 5	Kumma *et al.* (2008)
Soda Butte Creek, Montana and Cache Creek Wyoming, USA	In-channel hydrogeomorphic units including coarse woody debris	Airborne multi-spectral (blue, green, red and infrared bandwidths); 1 m resolution	Wright *et al.* (2000)
River Tummel, Scotland	Channel planform change and bank erosion	Aerial photography, 1:10,000 and 1:12,000	Winterbottom (2000)
North Pacific Rim rivers; North America and Russia	Juvenile salmonid habitat	Landsat TM; Quickbird and airborne multi spectral imagery	Whited *et al.* (2013)
(b) 3D and quasi-3D channel morphology and substrates			
North Ashburton River, New Zealand	Below water line morphology and exposed sediments	Aerial photography 1:3000	Westaway *et al.* (2000)
Waimakariri River, New Zealand	Channel morphology and DEM production in large braided systems	Aerial photography; 0.5 m resolution	Lane *et al.* (2000)
Platte River, Nebraska, USA	Channel bed topography	LiDAR	Kinzel *et al.* (2007)
Rhone River, France	Channel morphology and substrate	Drone and digital camera (5–14 cm resolution)	Lejot *et al.* (2007)
(c) 2D mapping of suspended solids concentrations and bed material			
River Ribble, northwest England	Percentage clay, silt and sand in inter-tidal sediments	Airborne thematic mapper data; 2 m pixel size	Rainey *et al.* (2000)
Saint Marguerite River, Canada	Gravel size	Gantry-mounted digital photography	Carbonneau *et al.* (2005)
Yangtze River, China	Suspended sediment	Landsat-7 ETM+	Wang *et al.* (2009)
(d) 2D and 3D mapping of floodplain morphology			
Madre de Dios, Peru	Floodplain geomophology and ecosystem structure	Landsat ETM+; JERS1; Radar C band	Hamilton *et al.* (2006)
River Tummel, Scotland	Hydromorphology	Airborne multi-spectral imagery; 1 m resolution (2 m resolution)	Gilvear *et al.* (2007)
(e) 2D mapping of flood inundation			
Amazon floodplain.	Mapping of the nature of inundation	JERS I	Alsdorf *et al.* (2007)
River Meuse, The Netherlands	Mapping of the extent of inundation	SAR	Bates and De Roo (2000)
(f) 2D and 3D mapping of overbank sedimentation, deposition and scour			
River Ob, Siberia	Floodplain deposition and scour	SAR	Smith and Alsdorf (1998)
Yuba River, California	Fllodpalin morphology and overbank sedimentation	Photogrammetry and aerial photography	Ghoshal *et al.* (2010)

epochs. More recently, airborne multispectral imagery has been used in attempts to map a wide-range of geomorphic features with mixed success (eg. Wright *et al.* 2000). The potential is high but increased knowledge of the spectral characteristics of geomorphic features is still required.

Channel planform change has also been the research focus of a number of fluvial geomorphologists and sequential sets of aerial photographs have commonly been used to detect change (eg Lapointe and Carson 1986; Werritty and Ferguson 1980) (Table 6.4). Large-scale changes in channel position are often apparent from visual interpretation, but geometric recitification is required for accurate measurement of change, especially for small changes in bank position. For example, although Gilvear *et al.* (1999) could identify large-scale changes in channel position on the Luangwa River, Zambia, visually (Fig. 6.9), geometric rectification and digitization of rivers bank lines within a GIS was necessary to detect changes in other channel features and to measure accurately bank erosion rates.

The accuracy of visual comparison, without the use of geometric rectification or rectification of photographs to each other using fixed ground-control points, will depend on the degree of tilt and distortion of the aerial photographs. Williams *et al.* (1979) managed to superimpose 39 photogrammetrically recovered bank profiles to measure retreat rate to an accuracy of 0.06 m per year for the Ottawa River, for which 16 aerial flights were available from 1921 to 1979. However, the temporal resolution of aerial photography will vary widely between regions. In many cases, particularly in the New World, the temporal resolution will be much less. If short-term change, particularly resulting from a single flood event, is of interest, commissioned flights are often necessary to gain temporal resolution (e.g. Winterbottom 2000). When comparing images, it is also necessary to remember that gross changes between dates will mask cycles of erosion and accretion, and perhaps channels ‘flipping’ position only to return to the original course at a later date.

Figure 6.8 An image of approximately 28 km by 21 km of the River Mississippi and its floodplain north of Vicksburg. The image was acquired in October 1994 by the spaceborne imaging radar C/X-band synthetic aperture imaging radar system (SIR-C/X-SAR).

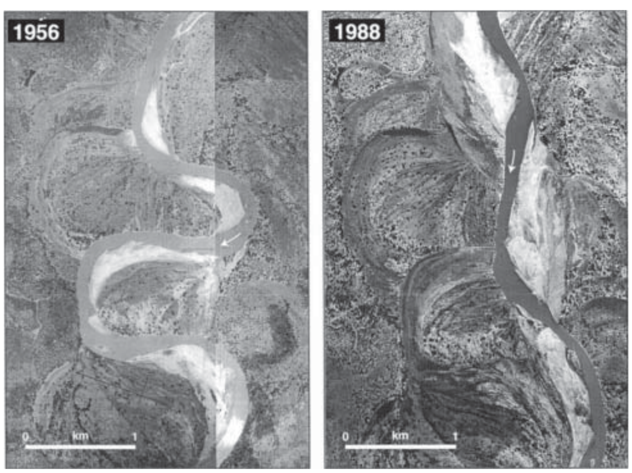


Figure 6.9 Channel change on the Luangwa River, Zambia, from 1956 to 1988. Source: Gilvear *et al.*, 1999. Reproduced with permission of Wiley.

Since the 1970s, satellite data have been used to enhance our knowledge of channel planform change of large river systems for which the synoptic view of a spaceborne sensor has advantages over the restricted coverage of individual aerial photographs (e.g. Phillip *et al.* 1989; Perez and Muller 1990). For example, Salo *et al.* (1986) used multi-date Landsat MSS images of the meandering and anastomosing stretches of the Ucayili and Amazon in Peru to quantify lateral migration rates of 200 m per year between 1979 and 1983. A more rudimentary approach based on the visual interpretation of photographic images was used by Ramasamy *et al.* (1991) to identify relict channels in the Yamuna River, Western India. Most remarkably, Jacobberger (1988) mapped abandoned river channels that were active in Sahelian Mali 6000 to 8000 years ago using MSS and TM images.

3D and quasi-3D channel morphology and channel change

Increasingly, various remote sensing methods have been used to produce 3D or quasi-3D reach-scale channel morphology. The major problem with quantifying channel morphology in three dimensions using remote sensing methods is the fact that a different approach is needed for exposed and submerged areas of the river channel. For exposed channel bars, large-scale aerial photogrammetry can be used to produce accurate elevation data (e.g. Westaway *et al.* 2000) (Table 6.4). Laser altimetry is also now being used to map the morphology of exposed channel beds, including mapping of change from sequential data sets (e.g. large braided rivers of New Zealand). Unfortunately, large areas of the channel are exposed under low-flow conditions only on some rivers. Hence a more important concern for the fluvial geomorphologist, and from a remote sensing perspective a more challenging problem, is that of mapping submerged areas.

A relatively robust technique for applying image analysis to aerial photographs to derive water depths for shallow non-turbid rivers has been developed in fairly recent years (Gilvear *et al.* 1995; Winterbottom and Gilvear 1997; Lane *et al.* 2000). The technique relies on a good correlation between the 'grey-tone' on aerial photographs and water depth. This is not always visible to the eye but with image enhancement can be detected. The variation in 'grey-tone' in shallow, clear water rivers and simple situations relates to variations in the reflectance of light from the river bed. The absorption of light radiation in water increases exponentially with depth and a number of algorithms have been developed to model this relationship (Lyzenga 1981; Clark *et al.* 1987; Bierworth *et al.* 1993) and to map the 3D morphology of gravel-bed rivers (e.g. Gilvear *et al.* 1995; Winterbottom 1995; Hicks *et al.* 1999). The results of these studies have proved to be relatively accurate in comparison with ground-truth data collected contemporaneously with imagery. Unless tied to direct measurements of the elevation of exposed sediments, this method produces only a quasi-3D model in that absolute elevations are not known.

Another approach to obtaining 3D channel morphology using combined remote sensing and ground data was that

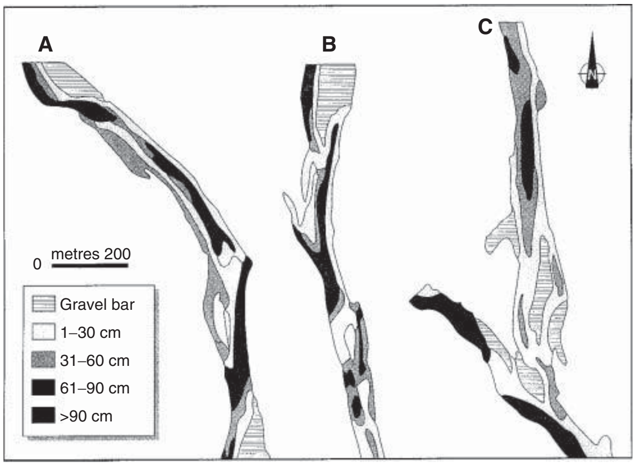


Figure 6.10 Example of a bathymetric map produced by image analysis applied ATM data of the confluence of the Rivers Tay and Tummel, Scotland. A, B and C are adjoining reaches in the downstream direction. Source: Winterbottom and Gilvear, 1997. Reproduced with permission of Wiley.

undertaken by Lane *et al.* (1994). Above water line topography was quantified repeatedly over a 21 day period by rigorous analytical photogrammetry applied to oblique aerial photography. Below water line measurements were undertaken using a rapid tacheometric survey and tied into the same ground-control network, producing daily 3D images of the stream bed. Differences between these images then formed the basis for calculation of bedload transport rates and zones of aggradation and degradation. For this case, the results suggest that a cross-section spacing of less than 2 m is required to estimate cut or fill on wide braided rivers (>10 m wide) to within 20% of the correct value. This demonstrates the need for rapid, high spatial resolution techniques for mapping 3D channel form to be developed for further understanding of channel bed dynamics.

Multi- and hyperspectral imagery can potentially provide the geomorphologist with the ability to detect variations in water depth in deeper and more turbid channels. Using airborne multispectral data, Winterbottom and Gilvear (1997) found the best relationship between water depth and radiance in the interval 605–625 nm (Fig. 6.10). Comparison of the 3D image produced by Winterbottom and Gilvear (1997) with a later image, produced by applying the same technique and which followed a 1 : 70 year return period flood, also allowed subtle changes in bedforms to be identified (Bryant and Gilvear 1999). In contrast to the work of Winterbottom and Gilvear (1997), Acornley *et al.* (1995), using CASI data, found that wavelengths of 800–820 nm produced the best results, although all bands in the ranges 510–610 and 645–820 nm gave satisfactory results. Here the imagery was captured in autumn when aquatic macrophytes were absent, which would have complicated image analysis. Using a multispectral video imaging system that could detect in the green (550 nm), red (650 nm) and near-infrared (850 nm) regions of the electromagnetic spectrum, Hardy *et al.* (1994) were also able to classify water depths and features such as runs, pools and riffles on the Green River, Utah.

Following these pioneering studies, further work has improved our knowledge of the effect of water column, water quality and bed type on depth measurements (Legleiter *et al.* 2004, 2009; Gilvear *et al.* 2007; Lejot *et al.* 2007; Feurer *et al.* 2008) and as such the technique is becoming more routinely incorporated into geomorphological investigations. The accuracy of such depth classifications is greatest in areas of low surface roughness, because a 'broken' water surface can scatter light and cause erroneous values; such phenomena, however, may aid the mapping of hydraulic habitat. Other limitations to the technique may include excessive shading of the bed, the presence of submerged, floating and emergent vegetation and high water turbidity. A range of new multispectral videography systems are also becoming available that should give good image geometry, a greater number of bands and spectral resolution and a convenience in deployment and data processing unmatched by more traditional systems (Sun and Anderson 1994; Hardy 1998). On larger rivers, bathymetric mapping has been undertaken using satellite data, but high turbidity and deeper water often preclude success. Vinod Kumar *et al.* (1997), however, undertook bathymetric mapping in the vicinity of the Rupnarayan–Hooghly river confluence, India, to depths of 8–10 m using LISS-II data in the wavelengths 0.77–0.80 nm. This was undertaken to guide dredging operations within the port of Calcutta.

Production of 3D images of bed topography for long channel reaches from remotely sensed data offers great potential for linking hydraulic modelling with channel change. The primary assumptions of these techniques are that both (i) the attenuation coefficient and (ii) the substrate reflectance remain constant over the full length and breadth of the extrapolated area. For the most part these assumptions will hold true for short river reaches, but ground data are needed to verify the assumptions or produce separate algorithms to account for differences in the attenuation coefficient and substrate.

Over the last decade, airborne and terrestrially mounted light imaging, detection and ranging (LiDAR) has also been added to the tools available to geomorphologists to map and detect change in 3D channel geometry. It is a device that is similar in operation to radar but emits pulsed laser light instead of microwaves. Each pixel that is scanned is assigned an x,y,z coordinate that allows for accurate 3D mapping of the object being scanned. Thoma and *et al.* (2005) advocated airborne laser scanning for riverbank erosion assessment and Kinzel *et al.* (2007) mapped the bed topography of shallow sand bed streams using an airborne-derived LiDAR dataset. McKean *et al.* (2008), using a narrow-beam, water-penetrating green LiDAR system (NASA's Experimental Advanced Airborne Research LiDAR – EARRL) were also able to map both floodplain and channel morphology over a 10 km reach on a mountain stream and link geomorphic patterns to salmon spawning (Fig. 6.11). Repeat LiDAR surveys of a river reach can also be used to determine accurately channel change and inferences made about sediment budgets (e.g. Fuller *et al.* 2003). Legleiter (2012)

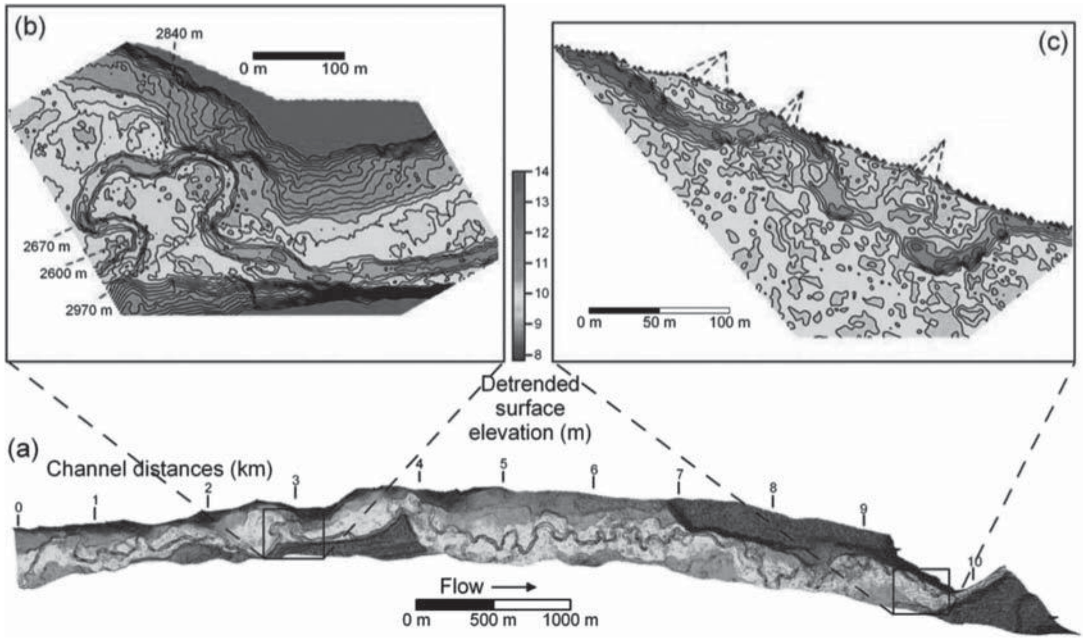


Figure 6.11 Channel and floodplain morphology for Bear Valley Creek, Idaho, USA, as determined by green LiDAR (EARLL) The digital terrain models are corrected for valley gradient. Source: McKean *et al.*, 2008. Reproduced with permission of ESA Journals.

has also developed a hybrid approach to bathymetric mapping. He fused bathymetric data from spectrally based methods with LiDAR data to provide maps that would not have been possible with just one of the methods.

2D mapping of turbidity, suspended solids concentrations and bed material

The use of satellite imagery on large river systems and airborne data on smaller rivers provides the opportunity to measure spatial and temporal changes in suspended sediment concentrations at the water surface over long reaches (Bustamante *et al.*, 2009; Wang *et al.*, 2009). However, because water chemistry, surface water roughness, sediment size, shape and mineralogy, atmospheric conditions and shadow all also affect the spectral properties of water, in addition to water depth in shallow areas, ground-truth data are often needed to allow calibration. Using a field-derived relationship between suspended sediment concentrations and spectral data, Aranuvachapun and Walling (1988) used satellite data to map spatial variability in suspended sediments for the Yellow River. The relationship between suspended sediment concentrations and reflectance has been explored more in the literature relating to coastal environments than for fluvial environments per se (e.g. Lathrop and Lillesand 1986; Novo *et al.* 1989a,1989b; Xia 1993; Ferrier 1995). Few studies have extended these analyses upstream to streams and rivers and problems may arise in shallow and heterogeneous aquatic environments where mixed pixels are also likely. However, the use of remotely sensed data in understanding the suspended sediment dynamics of medium-sized and large rivers is likely to increase.

Recent success has been obtained with mapping bed material size remotely (Table 6.4). This has been achieved using two different approaches. The first technique relies on the different spectral characteristics of sands, silts and gravels, mapping surificial sediment sizes on exposed inter-tidal areas (Rainey *et al.* 2000) (Fig. 6.12). In this situation, however, mapping was complicated by differences in soil moisture content because different areas were subject to differing drying times since the last high tide and spectral properties of sediments are moisture dependent (see earlier and Fig. 6.6). Image analysis techniques, which identify edges, have also been applied to high spatial resolution aerial photography to determine bed material size for river gravels, giving unparallel knowledge on patterns of grain size variability at multiple scales (Carbonneau, 2005; Carbonneau *et al.*, 2005).

6.4 Floodplain geomorphology and fluvial processes

2D and 3D mapping of floodplain morphology

Many landforms, including oxbow lakes, levees and scroll bars, are present on floodplains, resulting in a complex mosaic of topographical and sedimentological forms, often masked by vegetation. Identification of these features may be possible from variations in soil moisture and vegetation. Aerial photography has thus been used extensively to map floodplain features. Colour aerial photographs are particularly useful in that subtle differences in land cover that relate to underlying topography and sedimentology are more easily seen. Lewin and

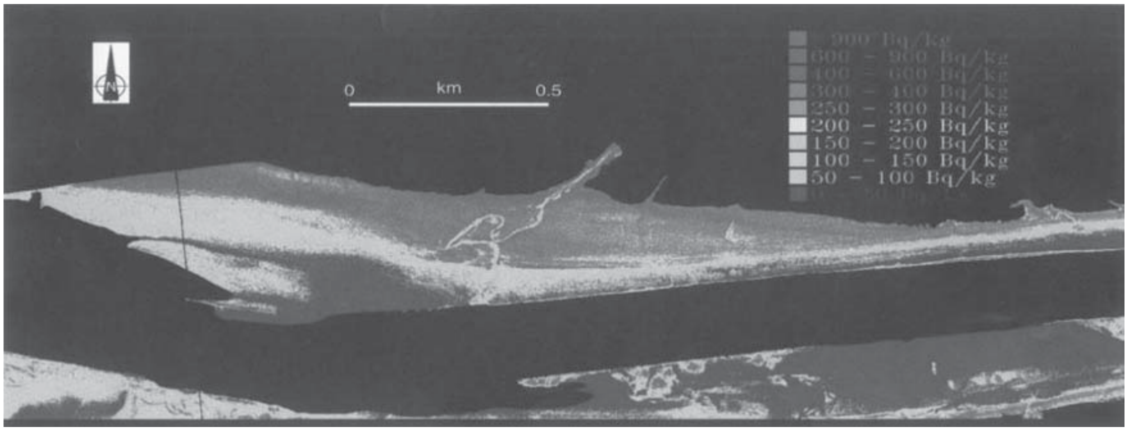


Figure 6.12 Particle size variation for the inter-tidal area of the River Ribble, England, as mapped using ATM data (Rainey 1999). Using a regression relationship between particle size and radionuclide concentrations on the river, the data were used to produce a map of radionuclide concentrations in bequerels per kilogram (Bq kg^{-1}), hence the key classification. Source: Rainey, 1999. Reproduced with permission of the author.

Manton (1975) used 1 : 5000 stereo pairs to map the floodplain topography of three Welsh rivers to a vertical resolution of 0.10 m and a horizontal resolution of 0.3 m. In the Garonne Valley, Muller (1992) found band 5 of TM imagery to be best for discriminating the floodplain from adjacent terraces and mapping spatial variability in alluvial surfaces on the floodplain. Davidson and Watson (1995) were able to map spatial variability in soil moisture on the floodplain. The areas of highest soil moisture were in topographic hollows left by relic channels. High spatial and vertical resolution topographic data can also be acquired by using scanning aircraft laser altimetry (Ritchie 1996). Laser altimetry is now being used by the UK Environment Agency to map floodplain topography for flood hazard mapping. Recent advances in the integration of scanning LiDAR technology with CCD digital imaging technology has produced airborne technology with access to real-time orthoimaging systems. The NASA ATM is a conically scanning airborne laser altimeter system capable of acquiring a swath width 250 m wide with a spot spacing of 1–3 m and a vertical precision of 10–15 cm. The potential of this in geomorphological and floodplain research has been demonstrated by Garvin and Williams (1993) and Marks and Bates (2000). Changes to the floodplain either side of a 1 : 65 year flood event were also quantified by Bryant and Gilvear (1999) using ATM data. Flood-induced depositional forms such as gravel lobes and sand splays were mapped. At a much larger scale, Trigg *et al.* (2012) mapped and classified all the stream systems on the middle reaches of the Amazon upstream of Manaus using ETM+ data (Fig. 6.13).

2D mapping of flood inundation

The use of airborne and satellite imagery to provide a synoptic perspective of flooding is relatively straightforward, except in forested floodplains, and has been extensively reviewed (e.g. Salomonson *et al.* 1983; Barton and Bathols 1989; Smith 1997). Sensor and platform use will depend upon the extent of inundation and spatial resolution, timing of the flood in relation to orbiting satellites or response times of airborne campaigns, the importance of emergent and floating vegetation and weather conditions. On small river systems, the extent of inundation can easily be seen on aerial photographs taken at the time of flooding. However, on such systems inundation is often short-lived and rarely are photographs available, particularly for the time of maximum inundation, which is often of greatest interest. Gilvear and Davies (unpublished work) were able to reconstruct the maximum extent of inundation using 1 : 5000 colour aerial photography taken 10 days after a 1 : 100 year flood event on the River Tay, Scotland, by the location of strand lines (i.e. flood debris).

When the flood coincides with a satellite orbit overhead and normally absence of clouds, and large areas of open water exist, inundation mapping can be undertaken simply using satellite imagery (Table 6.4). However, inundation mapping below a forest canopy can be problematic, although Ormsby *et al.* (1985) found that the L-band data from the Shuttle Imaging Radar (SIR-A) was helpful in separating forest vegetation from partially submerged grasses and shrubs and permitted a good definition of the land–water boundary even below a forest canopy. Cloud cover can be a problem in mapping inundation during the height of a flood except in the case of radar. The all-weather capability of radar is thus highly advantageous (Wagner 1994; Rudant 1994). Radar images record differences in roughness that indicate flood conditions. Sippel *et al.* (1994, 1998) used the scanning multichannel microwave radiometer on board the Nimbus 7 satellite to track changes in inundation on the Amazon River near Manaus over a 7 year period. Despite the coarse spatial resolution of the annual inundation area determined using mixing models, they correlated well with changes in river stage. Similarly, Brakenridge *et al.* (1998) were able to map the extent of flooding during the July 1993 flood on the Mississippi river using a SAR image of Iowa from the ERS-1 satellite. Moreover, by coupling SAR imagery with topographic imagery during the same flood, Brakenridge *et al.*

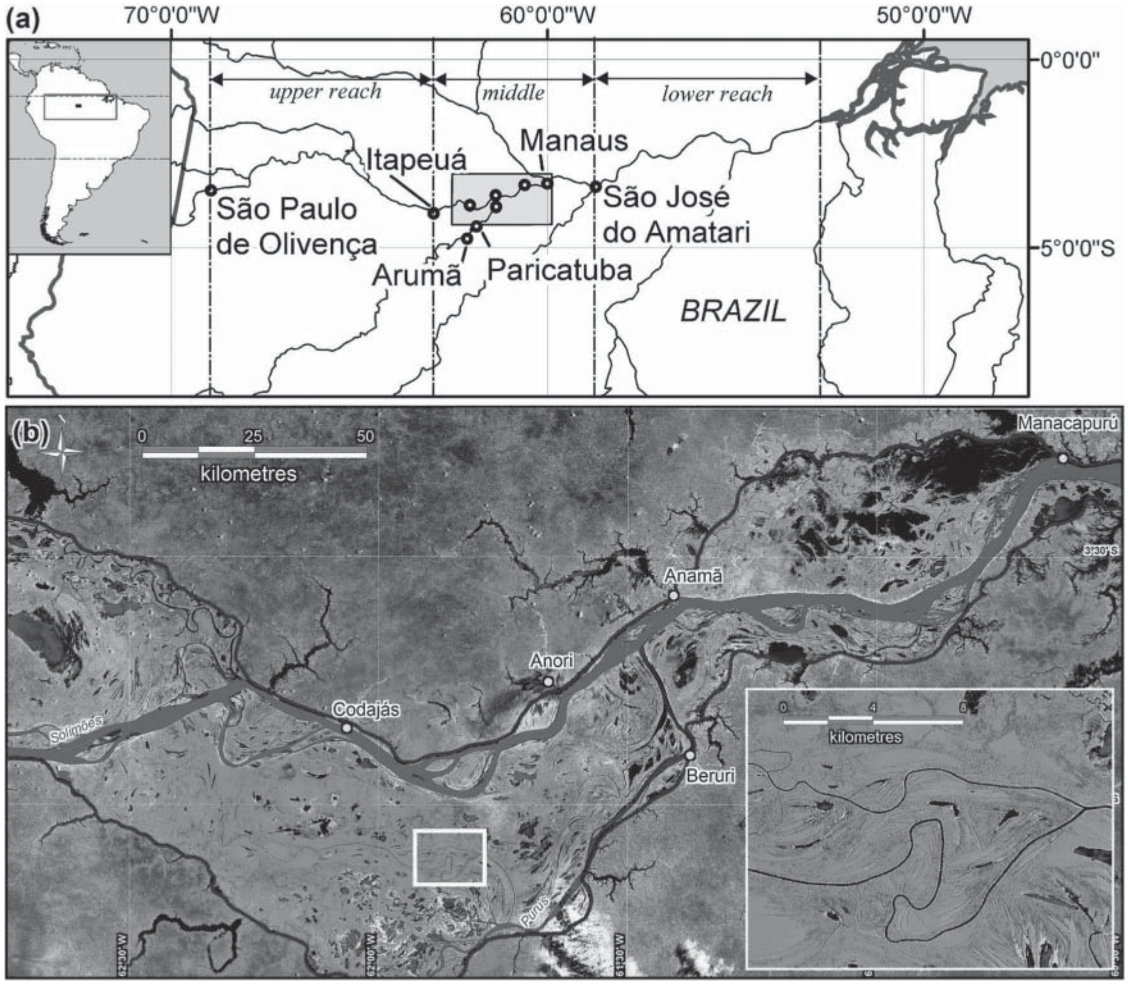


Figure 6.13 The location of a study by Trigg *et al.* (2012) (a) where they used an Orthorectified Pan-Sharpened ETM+ image mosaic (1999–2003) to digitize the stream network of the Amazon floodplain (b). Source: Trigg, 2012. Reproduced with permission of AGU Journals.

(1998) were able to measure water surface and hence the flood wave in Wisconsin. Alsdorf *et al.* (2007) were also able to demonstrate using remote sensing the hydraulic complexity of Amazonian flooding in time and space and as such increase flood inundation processes (Fig. 6.14).

2D and 3D mapping of overbank sedimentation, deposition and scour

Remotely sensed data also afford the possibility of deriving estimates of suspended sediment concentrations in flood waters and floodplain deposition. Mertes *et al.* (1993), working within the floodplain wetlands of the Amazon, showed that after nominal calibration to water–surface reflectance, near-surface suspended sediment concentrations could be estimated for each 30 m × 30 m pixel using linear spectral mixture analysis. Similarly, Gomez *et al.* (1995) used a Landsat 5 TM image to derive estimates of near-surface overbank suspended sediment concentrations in floodwaters during the 1993 Mississippi floods. Gomez *et al.* (1997), in conjunction with field measurements of deposition, were also able to use a TM image to produce a high spatial resolution map of floodplain sedimentation within the vicinity of the 1993 Sny Island levee break on the Mississippi near Canton, Missouri. Oblique aerial photography was also used to map scour, topsoil stripping, a sand rim and sand sheets close to the levee break, but spatial accuracy is compromised in such situations unless rigorous photogrammetric methods are adhered to. Evidence from field survey and 1 : 10,000 colour aerial photography (Gilvear and Black 1999) and ATM imagery (Bryant and Gilvear 1999), found similar geomorphological patterns to that of Gomez *et al.* (1997) in relation to flood embankment failures during a large flood in the same year on the River Tay, Scotland.

Synthetic aperture radar (SAR) interferometry also has some potential for assessing wide-scale floodplain erosion and deposition by allowing sequential construction of high-resolution digital elevation models (DEMs) and disturbance mapping from repeat-pass interferometric phase de-correlation. The latter is based on the fact that interferometric correlation or phase

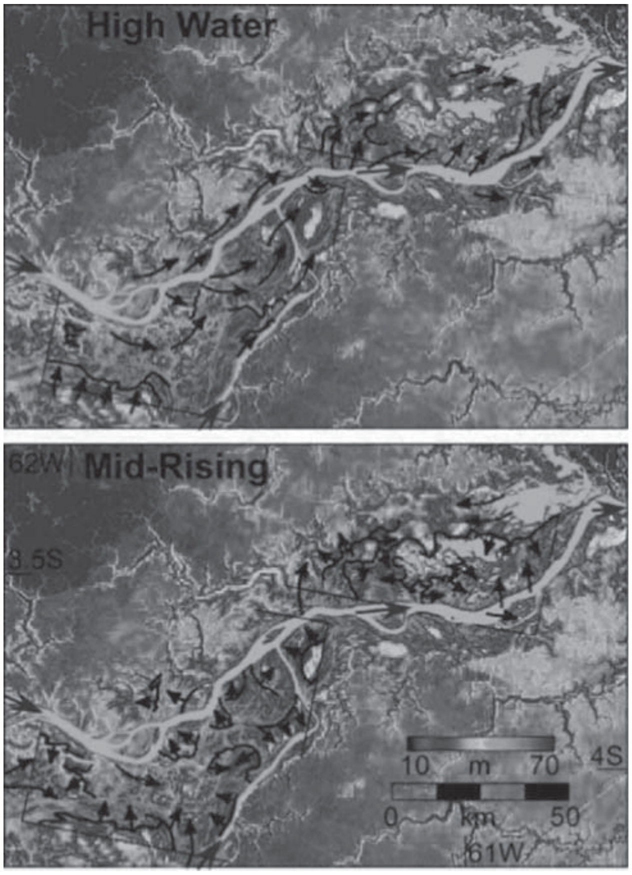


Figure 6.14 Patterns of inundation and flow direction on the Amazon floodplain during two flood conditions and derived from spaceborne interferometric synthetic aperture radar (SAR) JERS-1 measurements. Note the scale bar. Source: Alsdorf *et al.*, 2007. Reproduced with permission of AGU Journals.

coherence will decrease if the scattering properties of a surface change over time (Smith and Alsdorf 1998). Thus, floodplain scour or deposition can create a new scattering surface. While other factors can cause temporal phase de-correlation (e.g. soil moisture differences, vegetation growth), areas that do yield high phase coherence can be assumed to remain stable. Construction of accurate DEMs using SAR can also be problematic in heavily vegetated areas, but the method has been used successfully in identifying flood damage (Izenberg *et al.* 1996).

6.5 Conclusions

Analysis of aerial photography and remotely sensed data has wide application in detecting and mapping landforms, measuring temporal changes in fluvial landforms and controlling processes. The pros and cons of using differing sensor platforms and imagery are summarized in Table 6.5. For the study of large rivers (∼200 m wide or greater), spaceborne sensors provide the fluvial geomorphologist with information on channel morphology. For medium-sized rivers (∼20–200 m wide), data derived from airborne remote sensors or relatively large-scale aerial photography (approximately 1 : 5000 to 1 : 25,000) scale is better suited and can provide specific information. Increasingly, spaceborne systems will have the spatial resolution to map features on smaller rivers. Small river systems are more amenable to study by traditional terrestrial techniques and large-scale aerial photography (approximately 1 : 2500 or better) often taken from 'blimps' or remotely controlled aircraft. On small and medium-sized rivers, conventional photography with a hand-held camera can sometimes also be analysed to reveal information not otherwise obtainable at high spatial resolution.

Maximizing the potential of the analysis of aerial photographs and other remotely sensed data as a tool in the study of fluvial systems depends upon a sound understanding of the spatial and temporal capabilities of different sensors, the range and usefulness of a variety of image analysis techniques, the spectral characteristics of the fluvial environment and the nature and scale of the geomorphic problem under investigation. No one remote sensing system or type of image analysis provides the panacea, in that rivers vary in size, fluvial landforms and features have markedly different spectral characteristics and sensors vary in their spatial and spectral capability. Nevertheless, analysis of various types of terrestrially based and aerial photographs, data from first-generation satellite sensors and the latest generation of remote sensing systems offers the fluvial geomorphologist a rich set of tools. They allow the visualization, description and classification of a host of geomorphic attributes of rivers over a wide range of spatial scales and the detection and analysis of river channel change over time-scales from days to decades. Such information is crucial to planners interested in the stability of rivers before authorising adjacent developments, ecologists interested in fluvial disturbance and engineers concerned with river training or whether bridges, transport networks and flood defences may be threatened by erosion.

Acknowledgements

The UK Natural Environment Research Council (NERC) supported some of the investigations described here through the provision of airborne thematic mapper data and instrument loans from the NERC equipment pool for field spectroscopy. The authors' knowledge of the subject area has also been enhanced by collaborative projects with a number of individuals, including Dr Sandra Winterbottom, Dr Tim Malthus and Dr Ted Milton.

Table 6.5 Feasibility and advantages and disadvantages of various remote sensing approaches according to river channel size and type of geomorphic investigation (see also Table 6.3).

(A) Small sized river channels (<20 m wide)

Investigation type and remote sensing format	2D channel morphology and channel change	3D and quasi-3D channel morphology and channel change	2D mapping of suspended solids concentrations and bed material	2D and 3D mapping of floodplain morphology	2D mapping of flood inundation	2D and 3D mapping of overbank sedimentation, deposition and scour
Black-and-white aerial photography						
(General advantages) Widely available, relatively cheap and easy to commission flights which can coincide with cloudless skies. Historical record as far back as the 1940s and 1950s for most developed countries	*Advantages* 1:5000 scale or larger photography sufficient for detailed analysis of planform	*Advantages* Image analysis applied to scanned photographs can be used in clear, shallow streams to detect variations in water table depth		*Advantages* 1:5000 scale or larger photography sufficient for detailed mapping of floodplain landforms. Photogrammetry can be applied to stereo pairs for detection of relief	*Advantages* Feasible	*Advantages* 1:5000 scale or larger photography sufficient for detailed mapping of floodplain landforms. Photogrammetry can be applied to stereo pairs for detection of relief
	Disadvantages For long reaches (e.g. >2 km) a large number of individual photographs are needed, especially when the photography is not specially commissioned	*Disadvantages* Not possible in relatively deep or turbid rivers. Need for validation of water depths using ground-based measurements	*Disadvantages* Not generally possible unless very marked differences in concentrations are apparent. Need for ground-based measurements. Possibility of confusion with variations in stream bed reflectance in shallow streams	*Disadvantages* For long reaches a large number of individual photographs are needed especially when the photography is not specially commissioned. Experience in aerial photograph interpretation may be necessary for distinguishing some features	*Disadvantages* Confusion over classification of water with other land uses possible. For long reaches a large number of individual photographs are needed, especially when the photography is not specially commissioned	*Disadvantages* For long reaches a large number of individual photographs are needed, especially when the photography is not specially commissioned. Experience in aerial photograph interpretation may be necessary for distinguishing some features
Colour and infrared aerial photography						
(General advantages) As above but less widely available and limited historical record. Easier to interpret than above	As above	As above	As above but variability in turbidity often more marked than on black-and-white aerial photographs	As above although colour aerial photography can aid feature recognition. Infrared also enhances differences in soil moisture, which often aids feature recognition	As above but enhanced capability for water detection. On infrared photographs water shows up as black	As above although colour aerial photography can aid feature recognition. Infrared also enhances differences in soil moisture, which often aids feature recognition

(*continued overleaf*)

Table 6.5 (*continued*)

(A) Small sized river channels (<20 m wide)

Investigation type and remote sensing format	2D channel morphology and channel change	3D and quasi-3D channel morphology and channel change	2D mapping of suspended solids concentrations and bed material	2D and 3D mapping of floodplain morphology	2D mapping of flood inundation	2D and 3D mapping of overbank sedimentation, deposition and scour
Airborne multi- and hyperspectral imagery						
(General advantages) Digital format and specific spectral wavelengths can be prescribed according to the purpose of the study. Flights can be planned to coincide with cloudless skies	*Disadvantages* The scale or such imagery is not generally appropriate. Only on streams greater than 10 m wide would information be useful and then detection of bank position only accurate to approximately the nearest metre. Problems of mixed pixels	*Advantages* Enhanced water depth penetration at certain wavelengths. Laser altimetry provides the opportunity for accurate digital elevation models for exposed sediments (useful in braided rivers at low flows) *Disadvantages* The scale or such imagery is not generally appropriate. Problems of mixed pixels. Need for validation of water depths using ground-based measurements	*Advantages* Enhanced detection but confusion with bottom reflectance in non-turbid shallow stream *Disadvantages* The scale of such imagery is not generally appropriate. Problem of mixed pixels. Need for ground-based measurements	*Advantages* Laser altimetry provides the opportunity for accurate digital elevation models. Possible to map the floodplain surface through wooded canopies Use of specific wavebands aids feature recognition and soil moisture and particle size variations *Disadvantages* On small floodplains problems of spatial resoluton and mixed pixels	*Advantages* Easy detection of water surfaces. Possibility of estimating water depth	*Advantages* Use of specific wavebands aids feature recognition and soil moisture and particle size variations. Repeat flights using laser altimetry provide the opportunity for mapping areas that have undergone significant changes in erosion and deposition. Possible to map the floodplain surface through wooded canopies *Disadvantages* *On small floodplains problems of spatial resolution and mixed pixels*
Satellite and spaceborne imagery						
(General advantages) Widescale coverage and availability and high frequency of over flights. Spatial and spectral resolution becoming higher	Generally not appropriate except for crude channel planform detection	Not appropriate	Not appropriate	Generally not appropriate	Spatial resolution generally not appropriate except on small rivers with wide floodplains. Problems of mixed pixels	Generally not appropriate
Black-and-white aerial photography						
	Advantages 120,000 scale or larger photography sufficient for detailed analysis of planform	*Advantages* 1:10000 scale or larger photography sufficient. Image analysis applied to scanned photographs can be used in clear, shallow streams to detect variations in water table depth		*Advantages* 1:10000 scale or larger photography sufficient for detailed mapping of floodplain landforms. Photogrammetry can be applied to stereo pairs for detection of relief	*Advantages* Feasible using 1:20,000 scale or larger	*Advantages* 1:10000 scale or larger photography sufficient for detailed mapping of floodplain landforms. Photogrammetry can be applied to stereo pairs for detection of relief

Table 6.5 (*continued*)

(B) Medium-sized river channels (20–200 m wide)

River channel size/investigation type	2D mapping of channel morphology and channel change	3D mapping of channel morphology (3D)	2D mapping of suspended solids concentrations and bed material	2D and 3D mapping of floodplain morphology	2D mapping of flood inundation	2D and 3D mapping of overbank sedimentation, deposition and scour
	Disadvantages For reaches longer than 5 km more than one photograph is needed. Numbers of photographs may be large for long reaches, especially when the photography is not specifically commissioned	*Disadvantages* Not possible in relatively deep or turbid rivers. Need for validation of water depths using ground-based measurements. For reaches longer than 5 km more than one photograph is needed	*Disadvantages* Not generally possible unless very marked differences in concentrations are apparent. Need for ground-based measurements. Possibility of confusion with variations in stream bed reflectance in shallow streams	*Disadvantages* For long reaches a large number of individual photographs are needed, especially when the photography is not specially commissioned. Experience in aerial photograph interpretation may be necessary for distinguishing some features	*Disadvantages* Confusion over classification of water with other land uses possible. For long reaches a large number of individual photographs are needed, especially when the photography is not specially commissioned	*Disadvantages* For long reaches a large number of individual photographs are needed, especially when the photography is not specially commissioned. Experience in aerial photograph interpretation may be necessary for distinguishing some features
Colour and infrared aerial photography						
	As above	As above	As above but variability in turbidity often more marked than on black-and-white aerial photographs	As above although colour aerial photography can aid feature recognition. Infrared also enhances differences in soil moisture, which often aids feature recognition	As above but enhanced capability for water detection. On infrared photographs water shows up as black	As above although colour aerial photography can aid feature recognition. Infrared also enhances differences in soil moisture, which often aids feature recognition
Airborne multi- and hyperspectral imagery						
	Advantages Spatial resolution is often appropriate except at the lower end of this channel size range where mixed pixels are a problem and reduce the accuracy with which the position of river banks can be mapped	*Advantages* Spatial resolution is often appropriate except at the lower end of this channel size range where mixed pixels are a problem. Enhanced water depth penetration at certain wavelengths. Laser altimetry provides the opportunity for accurate digital elevation models for exposed sediments (useful in braided rivers at low flows)	*Advantages* Enhanced detection but confusion with bottom reflectance in non-turbid shallow streams	*Advantages* Laser altimetry provides the opportunity for accurate digital elevation models. Possible to map the floodplain surface through wooded canopies. Use of specific wavebands aids feature recognition and soil moisture and particle size variations	*Advantages* Easy detection of water surfaces. Possibility of estimating water depth and seeing water through wooded canopies	*Advantages* Use of specific wavebands aids feature recognition and soil moisture and particle size variations.Repeat flights using laser altimetry provide the opportunity for mapping areas that have undergone significant changes in erosion and deposition. Possible to map the floodplain surface through wooded canopies

(*continued overleaf*)

Table 6.5 (*continued*)

(B) Medium-sized river channels (20–200 m wide)						
River channel size/investigation type	2D mapping of channel morphology and channel change	3D mapping of channel morphology (3D)	2D mapping of suspended solids concentrations and bed material	2D and 3D mapping of floodplain morphology	2D mapping of flood inundation	2D and 3D mapping of overbank sedimentation, deposition and scour
		Disadvantages Need for validation of water depths using ground-based measurements. Not appropriate for depths greater than a couple of metres or turbid river systems	*Disadvantages* Need for ground-based measurements. Spatial resolution is often appropriate except at the lower end of this channel size range where mixed pixels are a problem	*Disadvantages* On small floodplains problems of spatial resolution and mixed pixels		*Disadvantages* *On small floodplains problems of spatial resolution and mixed pixels*
Satellite and spaceborne imagery						
	Generally not appropriate except for crude channel planform detection	Not appropriate	Not appropriate		Spatial resolution generally not appropriate except with small rivers with wide floodplains. Problems of mixed pixels	Generally not appropriate
Black-and-white aerial photography						
	Advantages 150,000 scale or larger photography sufficient for detailed analysis of planform	*Advantages* 1:10000 scale or larger photography sufficient. Image analysis applied to scanned photographs can be used in clear, shallow streams to detect variations in water table depth		*Advantages* 1:10000 scale or larger photography sufficient for detailed mapping of floodplain landforms. Photogrammetry can be applied to stereo pairs for detection of relief	*Advantages* Feasible using 1:20,000 scale or larger	*Advantages* 1:10000 scale or larger photography sufficient for detailed mapping of floodplain landforms. Photogrammetry can be applied to stereo pairs for detection of relief
	Disadvantages For reaches longer than 10 km more than one photograph is needed	*Disadvantages* Not possible in relatively deep or turbid rivers. Need for validation of water depths using ground-based measurements	*Disadvantages* Not generally possible unless very marked differences in concentrations are apparent. Need for ground-based measurements. Possibility of confusion with variations in stream bed reflectance in shallow streams	*Disadvantages* For long reaches a large number of individual photographs are needed, especially when the photography is not specially commissioned. Experience in aerial photograph interpretation may be necessary for distinguishing some features	*Disadvantages* Confusion over classification of water with other land uses possible. For long reaches a large number of individual photographs are needed, especially when the photography is not specially commissioned	*Disadvantages* For long reaches a large number of individual photographs are needed, especially when the photography is not specially commissioned. Experience in aerial photograph interpretation may be necessary for distinguishing some features

Table 6.5 (*continued*)

(B) Medium-sized river channels (20–200 m wide)

River channel size/investigation type	2D mapping of channel morphology and channel change	3D mapping of channel morphology (3D)	2D mapping of suspended solids concentrations and bed material	2D and 3D mapping of floodplain morphology	2D mapping of flood inundation	2D and 3D mapping of overbank sedimentation, deposition and scour
Colour and infra red aerial photography						
	As above	As above	As above but variability in turbidity often more marked than on black-and-white aerial photographs	As above although colour aerial photography can aid feature recognition. Infrared also enhances differences in soil moisture, which often aids feature recognition	As above but enhanced capability for water detection. On infrared photographs water shows up as black	As above although colour aerial photography can aid feature recognition. Infrared also enhances differences in soil moisture, which often aids feature recognition

(C) Large river channels (>200 m wide)

River channel size/investigation type	2D mapping of channel morphology and channel change	3D mapping of channel morphology (3D)	2D mapping of suspended solids concentrations and bed material	2D and 3D mapping of floodplain morphology	2D mapping of flood inundation	2D and 3D mapping of overbank sedimentation, deposition and scour
Airborne multi- and hyperspectral imagery						
	Scale is sufficient for overall planform detection and individual reaches can be zoomed in on for detailed analysis	*Advantages* Spatial resolution is not a problem Enhanced water depth penetration at certain wavelengths. Laser altimetry provides the opportunity for accurate digital elevation models for exposed sediments (useful in large braided rivers at low flows) *Disadvantages* Need for validation of water depths using ground-based measurements. Not appropriate for depths greater than a couple of metres or turbid river systems	*Advantages* Spatial resolution is not a problem. Enhanced detection but confusion with bottom reflectance in non-turbid shallow streams *Disadvantages* Need for ground-based measurements	*Advantages* Laser altimetry provides the opportunity for accurate digital elevation models. Possible to map the floodplain surface through wooded canopies Use of specific wavebands aids feature recognition and soil moisture and particle size variations	*Advantages* Easy detection of water surfaces. Possibility of estimating water depth and seeing water through wooded canopies	*Advantages* Use of specific wavebands aids feature recognition and soil moisture and particle size variations. Repeat flights using laser altimetry provide the opportunity for mapping areas that have undergone significant changes in erosion and deposition. Possible to map the floodplain surface through wooded canopies *Disadvantages* Huge volumes of data are generated even with modest channel lengths (e.g. 5 km)

(*continued overleaf*)

Table 6.5 (*continued*)

(C) Large river channels (>200 m wide)						
River channel size/investigation type	2D mapping of channel morphology and channel change	3D mapping of channel morphology (3D)	2D mapping of suspended solids concentrations and bed material	2D and 3D mapping of floodplain morphology	2D mapping of flood inundation	2D and 3D mapping of overbank sedimentation, deposition and scour
Satellite and spaceborne imagery						
	Advantages Spatial resolution generally appropriate except at the lower end of the size range and with lower resolution imagery	Water depths and turbidity usually too great	Ideal but need ground-based measurements for calibration	Ideal although mixed pixels cause detection of smaller features to be problematic. Use of specific wavebands aids feature recognition and soil moisture and particle size variations. Digital elevation modelling not normally feasible	Ideal.	Ideal although mixed pixels cause detection of smaller features to be problematic. Use of specific wavebands aids feature recognition and soil moisture and particle size variations. Sequential synthetic aperture radar interferometry allows disturbance mapping by identifying changes in the roughness of surfaces – difficult to use in well-vegetated areas
	Disadvantages More than one image for long river lengths (>100 km) needed					

References

Acornley, R.M., Cutler, M.E.J., Milton, E.J. and Sear, D.A. 1995. Detection and mapping salmonid spawning habitat in chalk streams using airborne remote sensing. In: *Proceedings of the 21st Annual Conference of the Remote Sensing Society, Southampton, September 1995*, pp. 267–274.

Alsdorf, D., Bates, P.D., Melack, J., Wilson, M. and Dunne, T. 2007. Spatial and temporal complexity of the Amazon flood measured from space. *Geophysical Research Letters* 34: L08402.

Aranuvachapun, S. and Walling, D.E. 1988. Landsat-MSS radiance as a measure of suspended sediment in the Lower Yellow River (Hwang Ho). *Remote Sensing of Environment* 25: 145–165.

Argialas, D.P., Lyon, J.G. and Mintzer, O.W. 1988. Quantitative description and classification of drainage patterns. *Photogrammetric Engineering and Remote Sensing* 54: 505–509.

Asrar, G. 1989. *Theory and Applications of Optical Remote Sensing*, New York: John Wiley & Sons, 725pp.

Barton, I.J. and Bathols, J.M. 1989. Monitoring floods with AVHRR. *Remote Sensing of Environment* 30: 27–35.

Bates, P.D. and De Roo, A.P.J. 2000. A simple raster-based model for flood inundation simulation, *Journal of Hydrology* 236(1–2): 54–77.

Brakenridge, G.R., Tracy, B.T. and Know, J.C. 1998. Orbital SAR remote sensing of a river flood wave. *International Journal of Remote Sensing* 19: 1439–1445.

Brivo, P.A., Colombo, R., Maggi, M. and Tomasoni, R. 2002. Integration of remote sensing data and GIS for accurate mapping of flooded areas. *International Journal of Remote Sensing* 23: 429–441.

Bryant, R.G. 1996. Validated linear mixture modelling of Landsat TM data for mapping evaporite minerals on a playa surface: methods and applications. *International Journal of Remote Sensing* 17: 315–330.

Bryant, R. and Gilvear, D.J. 1999. Quantifying geomorphic and riparian land cover changes either side of a large flood event using airborne remote sensing: River Tay, *Scotland, Geomorphology* 23: 1–15.

Bryant, R., Tyler, A, Gilvear, D., McDonald, P., Teasdale, I., Brown, J. and Ferrier, G. 1996. A preliminary investigation into the spectral characteristics of inter-tidal estuarine sediments. *International Journal of Remote Sensing* 17: 405–412.

Bustamante, J., Pacios, F., Diaz-Delgado, R. and Aragones, D. 2009. Predictive models of turbidity and water depth in the Donana marshes using Landsat TM and ETM+ images. *Journal of Environmental Management* 90: 2219–2225.

Carbonneau, P.E. 2005. The threshold effect of image resolution on image-based automated grain size mapping in fluvial environments. *Earth Surface Processes and Landforms* 30: 1687–1693.

Carbonneau, P.E. and Piégay, H., coord. 2012. *Remote Sensing and River Management*, Chichester: John Wiley & Sons.

Carbonneau, P.E., Bergeron, N.E. and Lane, S.N. 2005. Automated grain size measurements from airborne remote sensing for long profile measurements of fluvial grain sizes. *Water Resources Research* 41: W11426.

Chandler, J.H. 1999. Effective application of automated digital photogrammetry for geomorphological research, *Earth Surface Processes and Landforms* 24: 51–63.

Christensen, E.J., Jensen, J.R., Ramsey, E.W. and Mackey, H.E. Jr., 1988. Aircraft MSS data registration and vegetation classification for wetland change detection. *International Journal of Remote Sensing* 9: 23–38.

Clark, R.K., Fay, T.H. and Walker, C.L. 1987**.** Bathymetry calculations with Landsat 4 TM imagery under a generalised ratio assumption *Applied Opt*ics 26: 4036–4038.

Cosandier, D., Ivanco, T.A., Chapman, M.A. and Dylke, M. 1994. The integration of a digital elevation model in CASI image geocorrection. In: *Proceedings of the First International Airborne Remote Sensing Conference and Exhibition, Strasbourg, France*, Ann Arbor, MI: Environmental Research Institute of Michigan (ERIM), pp. 515–529.

Cracknell, A.P. 1998. Synergy in remote sensing – what's in a pixel? *International Journal of Remote Sensing* 19: 2025–2047.

Cracknell, A.P. and Heyes, L.W.B. 1993. *Introduction to Remote Sensing*, London: Taylor & Francis, 293pp.

Curran, P. 1994. Imaging spectrometry, *Progress in Physical Geography* 18: 247–266.

Davidson, D.A. and Watson, A. 1995. Spatial variability in soil moisture as predicted from airborne thematic mapper (ATM) data. *Earth Surface Processes and Landforms* 20: 219–230.

Dekker, T., Malthus, T. and Hoogenboom, P. 1995. The remote sensing of inland water quality. In: Danson, F.M. and Plummer, S.E., eds., *Advances in Environmental Remote Sensing*, London: Chapman & Hall, pp. 123–143.

Devereux, B.J., Fuller, R.M., Carter, L. and Parsell, R.J. 1990. Geometric correction of airborne scanner imagery by delauney triangles, *International Journal of Remote Sensing* 11: 2237–2251.

Diakite, M., Yerheau, M. and Bonn, F. 1986. The utility of Landsat TM imagery in the inland delta cartography of Mali. In: *Proceedings of the 20th International Symposium on Remote Sensing of Environment, Nairobi*, Ann Arbor, MI: Environmental Research Institute of Michigan (ERIM), pp. 567–574.

Dixon, L.F.J., Barker, R., Bray, M., Farres, P., Hooke, J., Inkpen, R., Merel, A., Payne, D. and Shelford, A. 1998. Analytical photogrammetry for geomorphological research. In: Lane, S., Richards, K. and Chandler, J., eds., *Landform Monitoring,* Modelling and Analysis, Chichester: John Wiley & Sons, pp. 63–94.

Ferrier, G. 1995. A field study of the variability in the suspended sediment concnetration–relectance relationship, *International Journal of Remote Sensing* 16: 2713–2720.

Feurer, D., Bailly, J.S., Peuch, C., Lecoarer, Y. and Viau, A. 2008. Very high resolution mapping of river immersed topography by remote sensing. *Progress in Physical Geography* 32: 1–17.

France, M.J., Collins, W.G. and Chindeley, T.R. 1986. Extraction of hydrological parameters from Landsat Thematic Mapper Imagery. In: *Proceedings of the 20th International Symposium on Remote Sensing of Environment, Nairobi*, Ann Arbor, MI: Environmental Research Institute of Michigan (ERIM), pp. 1165–1173.

Fuller, I.C., Large, A.R.G., Charlton, M.E., Heritage, G.L. and Milan, D.J. 2003. Reach-scale sediment transfers: an evaluation of two morphological budgeting approaches. *Earth Surface Processes and Landforms* 28: 889–903.

Gardner, T.W., Connors, K.F., Hu, H. and Peterson, G.W. 1987. Delineation of ephemeral fluvial networks on low-relief piedmont surfaces, Plutonium Valley, Nevada, USA. In: *Spot-1, Image Utilisation, Assessment Results*, Paris: CNES, pp. 215–220.

Garvin, J.B. and Williams, R. 1993. Geodetic airborne laser altimetry of Breidamerkurjokull and Skeidararjokull, Iceland and Jakobshavn, Greenland. *Annals of Glaciology* 17: 377–386.

Ghoshal, S., James, L.A., Singer, M.B. and Aalto, R. 2010. Channel and floodplain change analysis over a 100 year period: Lower Yuba River, Californa. *Remote Sensing* 27: 1797–1825.

Gilvear, D.J. and Black, A. 1999. Flood induced embankment failures on the River Tay: implications of climatically induced hydrological change in Scotland. *Hydrological Sciences Journal* 43: 1–16.

Gilvear, D.J., Waters, T. and Milner, A. 1995. Image analysis of aerial photography to quantify changes in channel morphology and in-stream habitat following placer mining in interior Alaska. *Freshwater Biology* 44: 101–111.

Gilvear, D.J., Waters, T. and Milner, A. 1998. Image analysis of aerial photography to quantify the effect of gold placer mining on channel morphology, interior Alaska. In: Lane, S., Richards, K. and Chandler, J., eds., *Landform Monitoring,* Modelling and Analysis, Chichester: John Wiley & Sons, pp. 195–216.

Gilvear, D.J., Winterbottom, S.J. and Sichingbula, H. 1999. Character of meander planform change on the Luangwa River, Zambia. *Earth Surface Processes and Landforms* 16: 1–24.

Gilvear, D.J., Hunter, P. and Higgins T. 2007. An experimental approach to the measurement of the effects of water depth and substrate on optical and near infra-red refelectance: a field-based asssessment of the feasibility of mapping submerged instream habitat. *International Journal of Remote Sensing* 28: 2241–2256.

Gomez, B., Mertes, L.A., Phillips, J.D., Milligan, F.J. and James, L.A. 1995. Sediment characteristics of an extreme flood: 1993 Upper Mississippi River valley. *Geology* 23: 963–966.

Gomez, B., Phillips, J.D., Milligan, F.J. and James, L.A. 1997. Floodplain sedimentation and sensitivity: summer 1993 flood, Upper Mississippi River valley. *Earth Surface Processes and Landforms* 22: 923–936.

Hamilton, S.K., Kellndorfer, J., Lehner, B. and Tobler, M. 2006. Remote sensing of floodplain geomorphology as a surrogate for biodiversity in a tropical river system (Madre de Dios, Peru). *Geomorphology* 89: 23–38.

Hardy, T.B. 1998. The future of habitat modelling and instream flow assessment techniques. *Regulated Rivers – Research and Management* 14: 405–420.

Hardy, T.B., Anderson, P.C., Neal, C.M.U. and Stevens, D.K. 1994. Application of multispectral videography for the deliniation of riverine depths and mesoscale hydraulic features. Paper presented at the American Water Resources Association meeting, Effects on Human Induced Changes on Hydrologic Systems, June 26–29, 1994, Jackson Hole, Wyoming.

Hickin, E.J. and Nanson, G.C. 1975. Lateral migration rates of meandering rivers. *Journal of Hydraulic Engineering* 110: 1557–1567.

Hicks, D.M., Lane, S.N. and Westaway, R.M. 1999. A metric eye in the sky: photogrammetric tools for riverbed surveys. *Water and Atmosphere* 7: 24–27.

Hooper, I.D. 1992. *Relationships Between Vegetation and Hydrogeomorphic Characteristics of British Riverine Environments: a Remotely Sensed Perspective*, unpublished PhD Thesis, University of Southampton.

Irons, J.R., Weismiller, R.A., and Peterson, G.W. 1989. Soil reflectance. In: Asrar, G., ed., *Theory and Applications of Optical Remote Sensing*, New York: John Wiley & Sons, pp. 336–428.

Izenberg, N.R., Arvidson, R.E., Bracket, R.A., Saatchi, S.S., Osburn, G.R. and Dohrenwend, J. 1996. Erosional and depositional patterns associated with the 1993 Missouri River floods inferred from SIR-C and TOPSAR radar data. *Journal of Geophysical Research* 101: 23149–23167.

Jacobberger, P.A. 1988. Mapping abandoned river channels in Mali through directional filtering of thematic mapper data. *Remote Sensing of the Environment* 26: 161–170.

Kaufman, Y, J. 1989. The atmospheric effect on remote sensing and its correction. In: Asrar, G., ed., *Theory and Applications of Optical Remote Sensing*, New York: John Wiley & Sons, pp. 189–211.

Kinzel, P.J., Wright, C.W., Nelson, J.M. and Burman, A.R. 2007. Evaluation of an experimental LiDAR for surveying a shallow, braided, sand-bedded river. *Journal of Hydraulic Engineering* 133: 838–842.

Kumma, M., Lu, X.X., Rasphone, A., Sarkkula, J. and Koponen, J. 2008. Riverbank changes along the Mekong river: remote sensing detection in the Vientiane–Nong Khai area. *Quaternary International* 186: 100–112.

Lane, S.N. 2000. The measurement of river channel morphology using digital photogrammetry. *Photogrammetric Record* 16(96): 937–957.

Lane, S.N., Richards, K.S. and Chandler, J.H. 1993. Developments in photogrammetry; the geomorphological potential. *Progress in Physical Geography* 17: 306–328.

Lane, S.N., Chandler, J.H. and Richards, K.S. 1994. Developments in monitoring and modelling small-scale river bed topography. *Earth Surface Processes and Landforms* 19: 349–368.

Lane, S.N., Westaway, R.M., Hicks, D.M. and Duncan, M.J. 2000. High resolution digital photogrammetry and image analysis for the measurement of large gravel-bed rivers. Paper presented at the Annual Conference of the Remote Sensing Society, University of Leicester, September 2000.

Lapointe, M.F. and Carson, M.A. 1986. Migration patterns of an asymmetric meandering river: the Rouge River, Quebec. *Water Resources Research* 22: 731–743.

Lathrop, R.G. and Lillesand, T.M. 1986. Use of thematic mapper data to assess water quality in Green Bay and Central Lake, Michigan. *Photogrammetric Engineering and Remote Sensing* 52: 349–354.

Legleiter, C.J. 2012. Remote measurement of river morphology via fusion of LiDAR topography and spectrally based bathymetry. *Earth Surface Processes and Landforms* 37: 499–518.

Legleiter, C.J., Roberts, D.A., Marcus, W.A. and Fonstad, M.A. 2004. Passive remote sensing of river channel morphology and instream habitat: physical basis and feasibility. *Remote Sensing of Environment* 93: 493–510.

Legleiter, C.J., Roberts, D.A. and Lawrence, R.L. 2009. Spectrally based remote sensing of river bathymetry. *Earth Surface Processes and Landforms* 34: 1039–1059.

Lejot, J., Delacourt, C., Piégay, H., Fourneier, T., Trémélo, M.-L. and Allemand, P. 2007. Very high spatial resolution imagery for channel bathymetry and topography from an unmanned mapping controlled platform. *Earth Surface Processes and Landforms* 32: 1705–1925.

Lewin, J. and Manton, M.M. 1975. Welsh floodplain studies: the nature of floodplain geometry. *Journal of Hydrology* 25: 37–50.

Lisle, R.J. 2006. Google Earth: a new geological resource. *Geology Today* 22, 29–32.

Lyzenga, D.R. 1981. Remote sensing of bottom reflectance and water attenuation parameters in shallow water using aircraft and Landsat data. *International Journal of Remote Sensing* 2: 71–82.

Marcus, W.A. and Fonstad, M.A. 2008.Optical remote mapping of rivers at sub meter resolutions and watershed extents. *Earth Surface Processes and Landforms* 33: 4–24.

Malthus, T.J., Place, C.J., Bennet, S. and North, S. 1995. An evaluation of the Airborne Thematic Mapper sensor for monitoring inland waters. Paper presented at the Remote Sensing Society Meeting, Southampton, 1995.

Marks, K. and Bates, P. 2000. Integration of high-resolution topographic data withfloodplain flow models. *Hydrological Processes* 14(11–12): 2109–2122.

McKean, J.A., Isakk, D.J. and Wright, C.W., 2008. Geomorphic controls on salmon nesting patterns described by a new narrow-beam terrestrial LiDAR. *Frontiers of Ecology and the Environment* 6: 125–130.

Mertes, L.A.K., Smith, M.O. and Adams, J.B. 1993. Estimating suspended sediment concentrations in surface waters of the Amazon River wetlands from Landsat images. *Remote Sensing of the Environment* 43: 281–301.

Milne, J.A. and Sear, D.A. 1997. Modelling river channel topography using GIS. *International Journal of GIS* 11: 99–519.

Milton, E.J., Gilvear, D.J. and Hooper, I.D. 1995. Investigating river channel changes using remotely sensed data. In: Gurnell, A. and Petts, G.E., eds., *Changing River Channels*. Chichester: John Wiley & Sons, pp. 277–301.

Miranda, F.P., Fonseca, L.E.N. and Carr, J.R. 1998. Semivariogram textual classification of JERS-1 SAR data obtained over a flooded area of the Amazon rainforest. *International Journal of Remote Sensing* 19: 549–556.

Muller, E. 1992. *Evaluation de la Bande TM5 pour la Cartographie Morpho-hydrogéologique de la Moyenne Vallée de la Garonne. Project SPOT4/MIR.* Paris: CNES.

Muller, E., Decamps, H. and Dobson, M.K. 1993. Contribution of space remote sensing to river studies. *Freshwater Biology* 29: 301–312.

Novo, E.M.M., Hansom, J.D. and Curran, P.J. 1989a. The effect of sediment type on the relationship between reflectance and suspended sediment concentration. *International Journal of Remote Sensing* 10: 1283–1289.

Novo, E.M.M., Hansom, J.D. and Curran, P.J. 1989b. The effect viewing geometry and wavelength on the relationship between reflectance and suspended sediment concentration. *International Journal of Remote Sensing* 10: 1357–1372.

Nykanen, D.K., Foufoula-Georgiou, E. and Sapozhnikov V.B. 1998. Study of spatial scaling in braided river patterns using synthetic aperture radar imagery. *Water Resources Research* 34: 1795–1807.

Ormsby, J.P., Blanchard, B.J. and Blanchard, A.J. 1985. Detection of lowland flooding using active microwave systems. *Photogrammetric Engineering and Remote Sensing* 51: 317–328.

Perez, J. and Muller, E. 1990. Suivi de l'évolution de la végétation riveraine du Rio Guanare (Venezuela) à l'aide de donées SPOT XS, LANDSAT MSS et de photos aériennes. *Troisièmes Journées Scientifiques du Réseau de Télédétection de l'UREF, 13–16 Novembre 1990, Toulouse*.

Phillip, G., Gupta, R.P. and Bhattacharya, A. 1989. Channel migration studies in the middle Ganga basin, India, using remote sensing data. *International Journal of Remote Sensing* 10: 1141–1149.

Plummer, S.E., Danson, F.M. and Wilson, A.K. 1995. Advances in remote sensing technology. In: Danson, F.M. and Plummer, S.E., eds., *Advances in Environmental Remote Sensing*, Chichester: John Wiley & Sons, pp. 1–8.

Ramasamy, S.M., Bakliwal, P.C. and Verma, R.P. 1991. Remote sensing and river migration in Western India. *International Journal of Remote Sensing* 12: 2597–2609.

Rainey, M.P. 1999. *Airborne Remote Sensing of Estuarine Intertidal Radionuclide Concentrations*, unpublished PhD Thesis, University of Stirling.

Rainey, M.P., Tyler, A.N., Bryant, R.G., Gilvear, D.J. and McDonald, P. 2000. The influence of surface and interstitial moisture on the spectral characteristics of inter-tidal sediments; implications for airborne image acquisition and processing *International Journal of Remote Sensing* 21: 3025–3038.

Ritchie, J.D. 1996. Airborne laser altimeters, remote sensing applications to hydrology. *Hydrological Sciences Journal* 41: 625–636.

Rudant, J.P. 1994. French Guiana through the clouds: first complete satellite coverage. *Earth Observation Quarterly* 44: 1–6.

Sabins, F.F. 1996. *Remote Sensing: Principles and Interpretation*, Sydney: Freeman, 298pp.

Salo, J., Kalliola, R., Hakkinen, I., Makinen, Y., Niemala, P., Puhakka, M. and Coley, P.D. 1986. River dynamics and the diversity of Amazon lowland forest. *Nature* 322: 254–258.

Salomonson, V.V., Jackson, T.J., *et al.* 1983. Water resources assessment. In: *Manual of Remote Sensing*, 2nd edn., Falls Church, VA: American Society of Photogrammetry, pp. 1497–1570.

Schumann, R.R. 1989. Morphology of Red Creek, Wyoming, an arid-region anastomosing channel system. *Earth Surface Processes and Landforms* 14: 277–288.

Sippel, S.J., Hamilton, S.K., Melack, J.M. and Choudhury, B.J. 1994. Determination of inundation area in the Amazon River floodplain using the SMMR 37 GHz polarization difference. *Remote Sensing of the Environment* 48: 70–76.

Sippel, S.J., Hamilton, S.K., Melack, J.M. and Novo, E.M.M. 1998. Passive microwave observations of inundation area and the area/stage relation in the Amazon River floodplain *International Journal of Remote Sensing* 19: 3055–3074.

Smith, L.C. 1997. Satellite remote sensing of river inundation area, stage and discharge: a review. *Hydrological Processes* 11(10): 1427–1439.

Smith, L.C. and Alsdorf, D.E. 1998. A control on sediment and organic carbon delivery to the Arctic Ocean revealed with satellite SAR: Ob River, Siberia. *Geology* 26: 395–398.

Sun, K. and Anderson, J.M. 1994. An easily deployable miniature, airborne imaging spectrometer, In: *Proceedings of the First International Airborne Remote Sensing Conference and Exhibition, Strasbourg, France*, Ann Arbor, MI: Environmental Research Institute of Michigan (ERIM), pp. 178–189.

Thoma, D.P., Gupta, S.C., Bauer, M.E. and Klirchoff, C.E. 2005. Airborne laser scanning for riverbank erosion assessment. *Remote Sensing of the Environment* 95: 493–501.

Thorne, C.R., Russell, A.P.G. and Alam, M.K. 1993. Planform pattern and channel evolution of the Brahmaputra River, Bangladesh. In: Best, J.L. and Bristow, C.S., eds., *Braided Rivers*, Geological Society Special Publication No. 75, London: Geological Society, pp. 257–277.

Townsend, P.A. and Walsh, S.J. 1998. Modelling floodplain inundation using an integrated GIS with radar and optical remote sensing. *Geomorphology* 21: 295–312.

Townshend, J.R.G. and Justice, C.O. 1988. Selecting the spatial resolution of satellite sensors required for global monitoring of land transformations. *International Journal of Remote Sensing* 9: 187–236.

Trigg, M.A., Bates, P.D., Wilson, M.D., Schumann, G. and Baugh, C. 2012. Floodplain channel morphology and networks of the middle Amazon River. *Water Resources Research* 58: 1–17.

Verstraete, M.M. and Pinty, B. 1992. Extracting surface properties from satellite data in the visible and near-infrared wavelengths. In: Mather, P.M., ed., *Terra-1: Understanding the Terrestrial Environment*, London: Taylor & Francis, pp. 203–209.

Vinod Kumar, K., Palit, A. and Bhan, S.K. 1997. Bathymetric mapping in Rupnarayan–Hooghly confluence using Indian remote sensing satellite data. *International Journal of Remote Sensing* 18: 2269–2270.

Wagner, M.J. 1994. ERS-1 images of the Christmas flood over Europe. *Earth Observation Quarterly* 43: 12–13.

Wang, J.J., Lu, X.X., Liew, S.C. and Zhou, Y. 2009. Retrieval of suspended sediment concentrations in large turbid rivers using Landsat ETM+: an example from the Yangtze River, China. *Earth Surface Processes and Landforms* 34: 1082–1092.

Warburton, J., Davies, T.R.H. and Mandl, M.G. 1993. A meso-scale field investigation of channel change and floodplain characteristics in an upland braided gravel bed river, New Zealand. In: Best, J.L. and Bristow, C.S., eds., *Braided Rivers*, Geological Society Special Publication No. 75, London: Geological Society, pp. 241–255.

Wealands, S., Stewardson, M., Gilvear, D., Hacker, J., Walker, J., Downes, B. and Rutherford, I. 2008. *Remote Sensing of Rivers: a Potential Contribution to the Murray–Darling Basin Sustainable Rivers Audit*, Melcourne: University of Melbourne.

Werritty, A. and Ferguson, R.I. 1980. Pattern changes in a Scottish braided river over 1, 30 and 200 years. In: Cullingford, R.A., Davidson, D.A. and Lewin, J., eds., *Timescales in Geomorphology*, Chichester: John Wiley & Sons, pp. 247–259.

Westaway, R.M., Lane, S.N. and Hicks, D.M. 2000. The development of an automated correction procedure for digital photogrammetry for the study of wide, shallow, gravel bed rivers. *Earth Surface Processes and Landforms* 25: 209–226.

Whited, D., Kimball, J. and Stanford, J. 2013. Estimation of juvenile salmon habitat in Pacific Rim rivers using multiscalar remote sensing and geospatial analysis. *River Research and Applications* 29: 135–148.

Williams, D.R., Romeril, P.M. and Mitchell, R.J. 1979. Riverbank erosion and recession in the Ottawa area. *Canadian Geotechnical Journal* 16: 641–650.

Wilson, A.K. 1997. An integrated data system for airborne remote sensing. *International Journal of Remote Sensing* 18: 1889–1901.

Winterbottom, S. 1995. *An Analysis of Channel Change on the Rivers Tay and Tummel, Scotland, Using GIS and Remote Sensing Techniques*, unpublished PhD Thesis, University of Stirling.

Winterbottom, S.J. 2000. Medium and short-term channel planform changes on the Rivers Tay and Tummel, Scotland. *Geomorphology* 34: 195–208.

Winterbottom, S.J. and Gilvear, D.J. 1997. Quantification of channel bed morphology in gravel-bed rivers using airborne multispectral imagery and aerial photography. *Regulated Rivers – Research and Management* 13: 489–499.

Wright, A., Marcus, W.M. and Aspinall, R. 2000. Evaluation of multi-spectral imagery as a tool for mapping stream morphology. *Geomorphology* 33: 107–120.

Xia, L. 1993. A united model for quantitative remote sensing of suspended sediment concentration. *International Journal of Remote Sensing* 14: 2665–2676.

CHAPTER 7

Geomorphic classification of rivers and streams

G. Mathias Kondolf[1], Hervé Piégay[2], Laurent Schmitt[3] and David R. Montgomery[4]
[1]University of California, Berkeley, CA, USA
[2]Université de Lyon, UMR 5600 CNRS, Lyon, France
[3]Université de Strasbourg, UMR 7362 CNRS, ENGEES, Strasbourg, France
[4]University of Washington, Seattle, WA, USA

You cannot step in the same river twice, for the second time it is not the same river.

– *Heraclitus*

7.1 Introduction

Rivers range widely in size, in channel form and in their degree of dynamism. Regional variability in river processes and river characteristics imparts a fundamental tension between attempts to develop generalizable classification systems and explicitly regional river approaches. It is not surprising, therefore, that attempts to classify rivers have resulted in a wide variety of classification schemes, serving a wide range of purposes from typologies for interpreting and understanding landscape evolution over geological time to those attempting to aid in the development of engineering designs for channel restoration projects. As with any tool, classification can be useful if applied properly to the appropriate problem. However, classification schemes are at best limited tools, whose capabilities are often overestimated by users lacking sound technical training in geomorphology. Moreover, reliance on classification systems can lead to serious problems, such as unnecessary and unwise interventions when misapplied or used in unskilled or inexperienced hands. This chapter discusses general philosophies of classifications in fluvial geomorphology, reviews examples of geomorphic classification systems and explores uses and limitations of classifications as a tool in fluvial geomorphology and river management.

Classification defined

Classification is the ordering of objects into groups based on common characteristics and attaching labels to the groups. Classification permits objects to be inventoried, so as to tally the number falling into various classes. If sub-groups of a collection of objects can be identified with common characteristics and behaviour patterns, distinct from other sub-groups, then a set of traits can be ascribed to the object (based on detailed study of other members of that class), which may then allow the prediction of the behaviour of the object under new circumstances. Classification may allow scientists to stratify an otherwise confusing universe into sets of similar objects, study representative objects and extrapolate results to other similar objects.

Classification refers both to the process of ordering objects in groups (the activity) and the resulting system of groups (the result). In common usage, the term is also used for the application of the resulting system, i.e. encountering new objects and placing them in the predetermined classes, a step referred to in the taxonomic literature as *identification* (Sneath and Sokal 1973). Taxonomists distinguish between *natural* classifications, a codification of natural clustering of objects with similar characteristics, and *special* classifications, which involve arbitrary distinctions drawn across a natural continuum (Sneath and Sokal 1973). Classifications of animals into species are considered natural classifications. Despite disagreements over details, most independent workers would reach similar classification decisions for major taxa, because evolution has provided a natural nested clustering. However, animals also can be organized into useful, albeit arbitrary, special classifications such as all carnivores or all aquatic invertebrates that cannot tolerate water temperatures exceeding a given level.

The process of classification development and application can be broken down into discrete steps. Based either on an *a priori* understanding of the system or cluster analysis on large data sets, a set of categories is proposed. The definition of categories depends, in part, upon the purpose of the classification, as a given set of objects can be classified in many different ways. The variables determined to be particularly diagnostic under the classification scheme are then emphasized in the subsequent collection of data. As additional objects are encountered, they are assigned to categories in the existing classification scheme (*identification*) or recognized as not fitting within

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

pre-existing categories. In the latter case, objects that do not fit into the classification can indicate the limits of the existing classification and thereby provide feedback for revising it and identifying a new group.

Classification and typology are often considered as synonymous, but are actually subtly distinct. Both terms focus on two aspects of the ordering process: (1) the clustering itself and (2) the variables on which the clustering is based. The typology is a classification of types, considered as features of a complex system, which can be considered a natural classification in the sense given by Sneath and Sokal (1973). In this chapter, we use the more general term ‘classification’. Buffington and Montgomery (2013, p. 730) emphasize the distinction between descriptive and process-based classifications, noting that descriptive classifications can be quantitative (based on measurement of physical parameters) and process-based classifications can be qualitative, but the key difference is that the latter are based on ‘mechanistic arguments and explanation of the physical processes associated with a given channel morphology’.

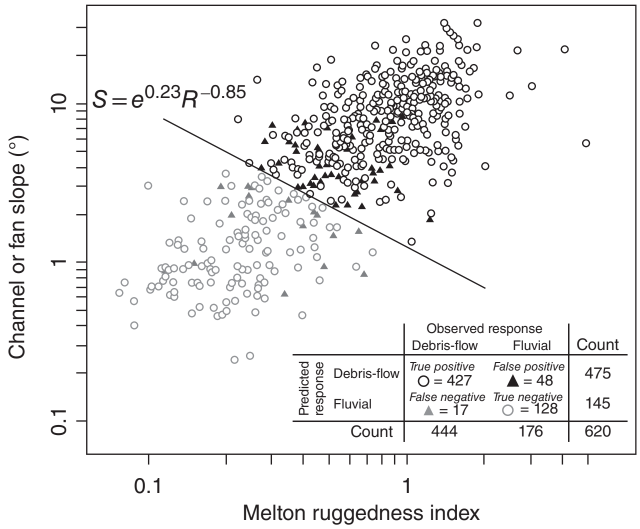


Figure 7.1 Small basins dominated by debris flows distinguished from those dominated by fluvial transport based on channel slope and the Melton Index. Source: Bertrand *et al*, 2013a. Reproduced with permission of Springer.

Purposes of classification

A wide range of classification schemes have been developed for fluvial systems, reflecting the intended purpose of the classification, different disciplines involved and the characteristics of the systems being classified (i.e. the studied environment, as per Gurnell *et al.* 1994). Classification can focus on spatial features such as river patterns, floodplains, in-channel features (e.g. pools and riffles), which can be separated according to a set of parameters, some of them being descriptive of the form itself (width, depth, slope, length, geometry of nested features such as braided index for a braided reach) or inferring differences in terms of functioning linking forms and associated processes. For example, Buffington and Montgomery (2013) illustrated this aspect showing hillslope stability according to valley side and channel gradients or risk of a side-slope mass wasting event entering the channel as a function of channel width relative to valley width. Bertrand *et al.* (2013a) showed that it is possible to distinguish small basins dominated by debris flows from those dominated by fluvial sediment transport, based on two variables, the Melton index and the channel or fan slope (Fig. 7.1).

We can distinguish two main objectives for river classification: (1) scientific understanding of how rivers function (e.g. existence of natural thresholds that produce longitudinal complexity on different spatial scales and whether channels can be clustered in ‘homogeneous’ classes) and (2) geomorphically based management guidance to inform decisions about channel maintenance, improvement, restoration or conservation. In the latter case, geomorphic criteria may be complemented by criteria from other disciplines (e.g. ecology, water chemistry). These two kinds of objectives are not mutually exclusive but can be linked through elaborating river classifications based on quantitative and qualitative field studies that integrate in a hierarchical view that distinguishes between independent and dependent variables, forms, processes, channel adjustments (Brookes 1987; Downs 1994) and temporal trajectories (Brierley and Fryirs 2005). Although such classifications can potentially serve several uses in physical habitat assessment, restoration and evaluating the impacts of engineering works (Malavoi 2000), they are easier to elaborate in relatively small catchments.

A fundamental motivation for using classification of forms is to simplify complexity and improve communication and understanding, especially in interdisciplinary settings. As biologists sought to classify aquatic habitat components to provide a common framework for the input of diverse disciplines and sites (Platts 1980; Hawkins *et al.* 1993), a number of authors recognized the need for a classification system for stream channels (Pennak 1971; Hawkes 1975; Terrell and McConnell 1978; Newson and Newson 2000). Classification systems intended to improve communication should be objective, so that operators from different disciplines and in different regions will reach the same classification decisions.

Another aim of classifications is more practical, namely to identify spatial units on which various management policies are implemented. For example, Bavarian Water Law uses a classification system to assign responsibility for river maintenance and flood control of large rivers (Class 1) to the state, medium-sized rivers (Class 2) to the seven districts within Bavaria and smaller watercourses (Class 3) to local communities (W. Binder, Bayerisches Landesamt für Wasserwirtschaft, Munich, Germany, personal communication, 1991). The classification system serves admirably for this administrative purpose, but it may not serve for other purposes, such as distinguishing among rivers with different ecological characteristics. Similarly, the US Forest Service incorporates fishery and water supply values in a classification system used to determine the degree of protection from timber harvest impacts afforded to a reach. The system is based on presence of perennial flow, presence of resident or anadromous fish, use for municipal water supply and relative size of stream (Gregory and Ashkenas 1990).

The influence of the classifier's discipline is readily apparent in the diverse classification systems proposed for South African rivers: geographic, limnological, chemical and biological criteria result in different groupings (King *et al.* 1992). Classification systems based on variables relevant to different disciplines can produce entirely different groupings, such as those based on bed-material size (ASCE 1992), water quality (BES 1990), macrophytes (Holmes 1989), invertebrates and fish (Pennak 1971; Furse *et al.* 1984), recreational potential (Zachman 1984), restoration potential (NRA 1992) or stability characteristics for engineering works (Simons 1978).

Finally, the geomorphic characteristics of fluvial systems in the region under study influence the resultant classification. For example, many channel classes used by the Tsongas National Forest in southeast Alaska, such as 'beaver dam/pond' channels and 'deeply incised muskeg' channels (Paustian *et al.* 1992) would have little relevance in England, where basin lithology (e.g. chalk or clay) is a principal determinant of channel form (Holmes 1989), or in the Great Plains of North America, where the percentage of silt and clay in the river bed and banks is a good predictor of channel processes and morphology (Schumm 1963). Table 7.1 summarizes many classifications based on geomorphic criteria, organized by objectives.

Classifications vary in spatial scale. Those that focus on process understanding may address spatial context only secondarily, sampling spatial units but not always mapping them.

Table 7.1 Examples of geomorphic-based river classification and sectorization objectives.

Objective	Scales	References
Describe valley geomorphology, quantify drainage network	Basin, valley, drainage network	Davis 1899; Strahler 1957
Classify and characterize hydrologic regimes	Basin	Gustard 1992
Provide a theoretic hierarchical framework for river classification	All scales	Hynes 1975; Schumm 1977; Lotspeich 1980; Brussock *et al.* 1985; Frissell *et al.* 1986; Kern 1994
Elaborate hierarchical typologies and/or ecoregional studies	All scales	Rohm *et al.* 1987; Cupp 1989a; Hugues *et al.* 1993; Omernik 1987; Wasson *et al.* 1993; Imhof *et al.* 1996; Allan and Johnson 1997; Heritage *et al.* 1997; Souchon *et al.* 2000
Characterize valley bottom or floodplain dynamics	Valley bottom, floodplain	Galay *et al.* 1973; Cupp 1989b; Nanson and Croke 1992; Bravard and Peiry 1999; Ferguson and Brierley 1999
Describe (or predict) alluvial channel patterns	Channel pattern	Leopold and Wolman 1957; Galay *et al.* 1973; Rust 1978; Schumm 1985; Paustian *et al.* 1984; Van den Berg 1995; Nanson and Knighton 1996; Alabyan and Chalov 1998
Regionalize channel morphology and dynamic	Channel reach, often viewed in the basin context	Petit 1995; Rosgen 1996
Sectorize streams in reach having homogeneous geomorphic functioning for management purposes	Channel reach, often viewed in the basin context	Maire and Wilms 1984; Cupp 1989b; Agence de l'Eau Rhin-Meuse *et al.* 1991; Orlowski *et al.* 1995; Van Niekerk *et al.* 1995; Bernot *et al.* 1996; Heritage *et al.* 1997; Schmitt 2001
Classify streams for management purposes	Channel reach, often viewed in the basin context	NRA 1993; Corbonnois and Zumstein 1994; Rosgen 1994, 1996; Zumstein and Goetghebeur 1994; Bernot and Creuzé des Châtelliers 1998; Doyle *et al.* 2000; Schmitt 2001; Piégay *et al.* 2009; Belletti *et al.* 2013
Classify streams on the basis of their morphodynamic processes and adjustments	Channel reach, often viewed in the basin context	Kellerhals, *et al.* 1976; Schumm 1963, 1977; Tricart 1977; Brookes 1987; Whiting and Bradley 1993; Downs 1994, 1995; Montgomery and Buffington 1997; Schmitt 2001; Emery *et al.* 2003; Orr *et al.* 2008
Classify reference *natural* states of streams (*Leitbild*; German approaches)	Channel reach, often viewed in the basin context	Otto and Braukmann 1983; Otto 1991; Müller *et al.* 1996; Bostelmann *et al.* 1998a,1998b; Tölk 1998
Identify reaches sensitive to erosion	Channel reach	Piégay *et al.* 1997
Identify reaches producing/storing LWD	Channel reach	Piégay *et al.* 1996
Stratify a River Quality Index	Channel reach, often viewed in the basin context	AQUASCOP 1997; Raven *et al.* 1997; Malavoi 2000; Schmitt 2001
Identify reaches for rehabilitation purposes	Channel reach, often viewed in the basin context	NRA 1992; Bostelmann *et al.* 1998a,1998b; Brierley and Fryirs 2000
Manage biological resources	All scales	Otto and Braukmann 1983; Wright *et al.* 1984; Cupp 1989a; Biggs *et al.* 1990; Souchon *et al.* 2000
Identify aquatic habitats/make biotic typologies (fish, macro-invertebrate, macrophytes)	Channel reach, morphodynamic unit and microhabitat, often viewed in the basin context	Huet 1949; Pennak 1971; Vannote *et al.* 1980; Wright *et al.* 1984; Cupp 1989a; Holmes 1989; Malavoi 1989; Biggs *et al.* 1990; Hawkins *et al.* 1993; Robach *et al.* 1996; Allan and Johnson 1997; Nicolas and Pont 1997; Montgomery *et al.* 1998; Beechie *et al.* 2005; Harvey *et al.* 2008

The main aim is to understand differences between groups and why, as manifest in clusters of points on graphs. Classifications that focus on an explicit spatial context can explore spatial organization, as is often done for management purposes, where fluvial forms with given characteristics are located to design a targeting or planning policy. Such a spatially explicit approach can be viewed on a river continuum, such as longitudinal zonation. Stream reaches and classes can be defined based on the downstream variation in stream power (Knighton 1999), using discontinuities in specific stream power to draw boundaries between reaches (Bernot *et al.* 1996; Astrade and Bravard 1999; Schmitt *et al.* 2001, 2007; Vocal Ferencevic and Ashmore 2012). However, classification can be also applied at a regional scale, discretizing the entire stream network, introducing regional factors such as lithology, hydroclimatic setting and river history. One of the main challenges is to integrate site-scale observations with network-scale classes, which can be facilitated by using GIS and remote sensing information (Alber and Piégay 2011).

Hierarchy in fluvial geomorphic classification

Fluvial systems can be viewed as inherently hierarchical, with smaller forms nested within larger ones (Fig. 7.2). In decreasing scale these could include landscape/ecoregion, floodplain/corridor (valley segments), channel reach and specific channel units (e.g. pools and riffles) and microhabitats (Lotspeich 1980; Amoros *et al.* 1982; Frissell *et al.* 1986) (Fig. 7.3). Lower hierarchical levels are controlled asymmetrically by the upper levels, i.e. upper levels control lower levels but not vice-versa (Naiman *et al.* 1992; Amoros and Petts 1993), because within a given landscape ecoregion, similar lithology, climate, geomorphology and land-use history would tend to give rise to similar stream characteristics and thus constitute a stream system class, within which classes could be defined for progressively smaller features. The asymmetric control of small-scale features by larger-scale characteristics implies that one must see beyond local site conditions to understand catchment controls, to view streams in a watershed context (Hynes 1975; Frissell *et al.* 1986). Moreover, it implies that stream classes developed for one region need not be applicable elsewhere. Also, at the spatial scale of valley segments (floodplain/corridor), inherited geomorphological features such as coarse fluvio-glacial deposits and the legacy of fluvial palaeodynamics can have a considerable influence on present channel dynamics and adjustment potential (Brierley and Fryirs 2005; Schmitt *et al.* 2007).

Because rivers typically undergo profound changes along their length, each level of a classification system must either limit itself to homogeneous sections of channel (Kellerhals *et al.* 1976; Brice 1982; Rosgen 1994; Montgomery and Buffington 1997; Montgomery *et al.* 1998) or address the nature of longitudinal change as a basis for classifying different regions (Brussock *et al.* 1985; Frissell *et al.* 1986; Bethemont *et al.* 1996; Montgomery 1999; Schmitt *et al.* 2007).

Underlying philosophies: rivers as a continuum or discrete types

One interesting aspect of the wide range of views on classification is an often unstated difference in underlying theoretical framework. As applied to river classification, the issue boils down to whether river systems are composed of a continuum of channel morphology or discrete types of channels either bounded by geomorphic thresholds or controlled by local influences such as a flow constriction imposed by a landslide deposit, differences in bed or clast lithology, confluences or changes of valley bottom width. In the latter case, it may be possible to develop a natural classification, whereas in the former case, all channel classification schemes are perforce arbitrary, special classifications, as was concluded from attempts to define

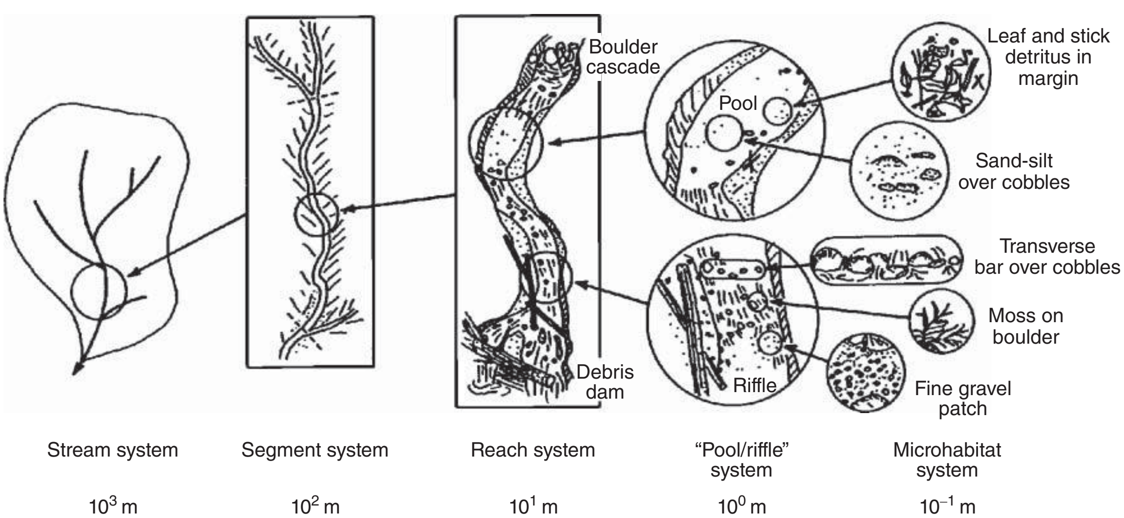


Figure 7.2 Hierarchical organization of a stream system and its habitat subsystems for second- or third-order mountain streams. Source: Frissell, 1986. Reproduced with permission of Elsevier.

	Variables										Spatial scales (km^2)					
	Geographical unit (geology, landform...)	Valley or valley bottom morphology	Channel pattern, sinuosity	Channel morphology	Gradient	Stream power	Channel bottom grain size	Sediment load (type and/or intensity)	Morphodynamic units	Morphodynamic adjustments	10^3 and more	10^2	10^1	10^0	10^{-1}	10^{-2}
Leopold and Wolman (1957)			●										●	●	●	
Rust (1978)			●										●	●	●	
Lotspeich (1980)	●										●	●	●	●		
Cloots and Maire (1980)	●	●	●	●								●	●	●		
Ferguson (1981)			●			●							●	●		
Maire and Wilms (1984)		●	●	●	●		●			●				●	●	
Brussock *et al*. (1985)	●										●					
Frissel *et al*. (1986)	●	●	●	●	●		●	●	●	●	●	●	●	●	●	●
Hugues *et al*. (1993)	●										●					
Cupp (1989a, 1989b)		●	●										●	●	●	
Agence de l'Eau R-M *et al*. (1991)	●			●	●								●	●	●	
Otto (1991)		●	●		●		●						●	●	●	
Nanson and Croke (1992)		●	●	•		●	●	●		●			●	●		
Corbonnois and Zumstein (1994)	●	●			●								●	●		
Downs (1994, 1995)	●				●					●			●	●	●	
Rosgen (1994, 1996)		●	●	●	●		●							●	●	●
Petit (1995)			●			●				●			●	●		
Nanson and Knighton (1996)	●	●	●	●	●	●	●	●		●			●	●		
Bethemont *et al*. (1996)	●	●	●	●	●				●		●	●	●	●	●	●
Bernot *et al*. (1996)	●			●	●	●	●						●	●		
AQUASCOP (1997)	●	●	•	•	•	●	•	●				●	●			

• : Variable partially taken into account

Figure 7.3 Variables taken into account and spatial scales for 21 geomorphic channel classification schemes.

objectively discrete classes from large data sets with respect to ecology (Cushing *et al.* 1983) and water quality (Wright *et al.* 1984).

Recognizing that fluvial forms vary in a longitudinal direction, longitudinal zonations for rivers (from headwaters to the sea) have been proposed for New Zealand rivers (Nevins 1965) and for Washington state (Palmer 1976). Both of these approaches identified four 'geo-hydraulic river zones', each with distinct channel gradient, channel pattern, valley cross-section, bed material size and 'material budget' (whether the bed is eroding, depositing or stable) (Fig. 7.4). Likewise, Schumm (1977) proposed the concept of fluvial systems (see Chapter 5) involving a general division of river systems into erosional headwater reaches, connected by transport reaches, ending in depositional zones. Analogous longitudinal zonations in biological characteristics have been proposed in general or for other regions (Carpenter 1928; Huet 1949; Vannote *et al.* 1980).

Local thresholds are important in explaining the complex succession of fluvial forms along the river system and the general zonations described above become complicated in rivers with complex geology and anthropic modifications. On the Ubaye River, an Alpine tributary to the Durance River in southern France, the longitudinal succession of channel form does not follow the conventional pattern because of its particular geological setting: a braided pattern occurs upstream in a wide marly basin, whereas downstream it transitions from wandering to meandering and then to straight, with increasing slope and grain size and decreasing valley bottom width as it traverses more competent lithologies (Piégay *et al.* 2000).

River channels exhibit characteristics of both a downstream continuum and locally controlled systems (Montgomery 1999). For example, channels generally widen downstream as a power function of drainage area in both alluvial and bedrock channels (Leopold and Maddock 1953), but local differences in lithology can affect the width of bedrock channels (Montgomery and Gran 2001). Valley morphology typically shows this complex intertwining of local controls and downstream continuum (Schmitt *et al.* 2007; Notebaert and Piégay 2013). Bed material gradually changes from cobbles and boulders in steep mountain channels to sand and gravel in lowland rivers, but local tributary inputs

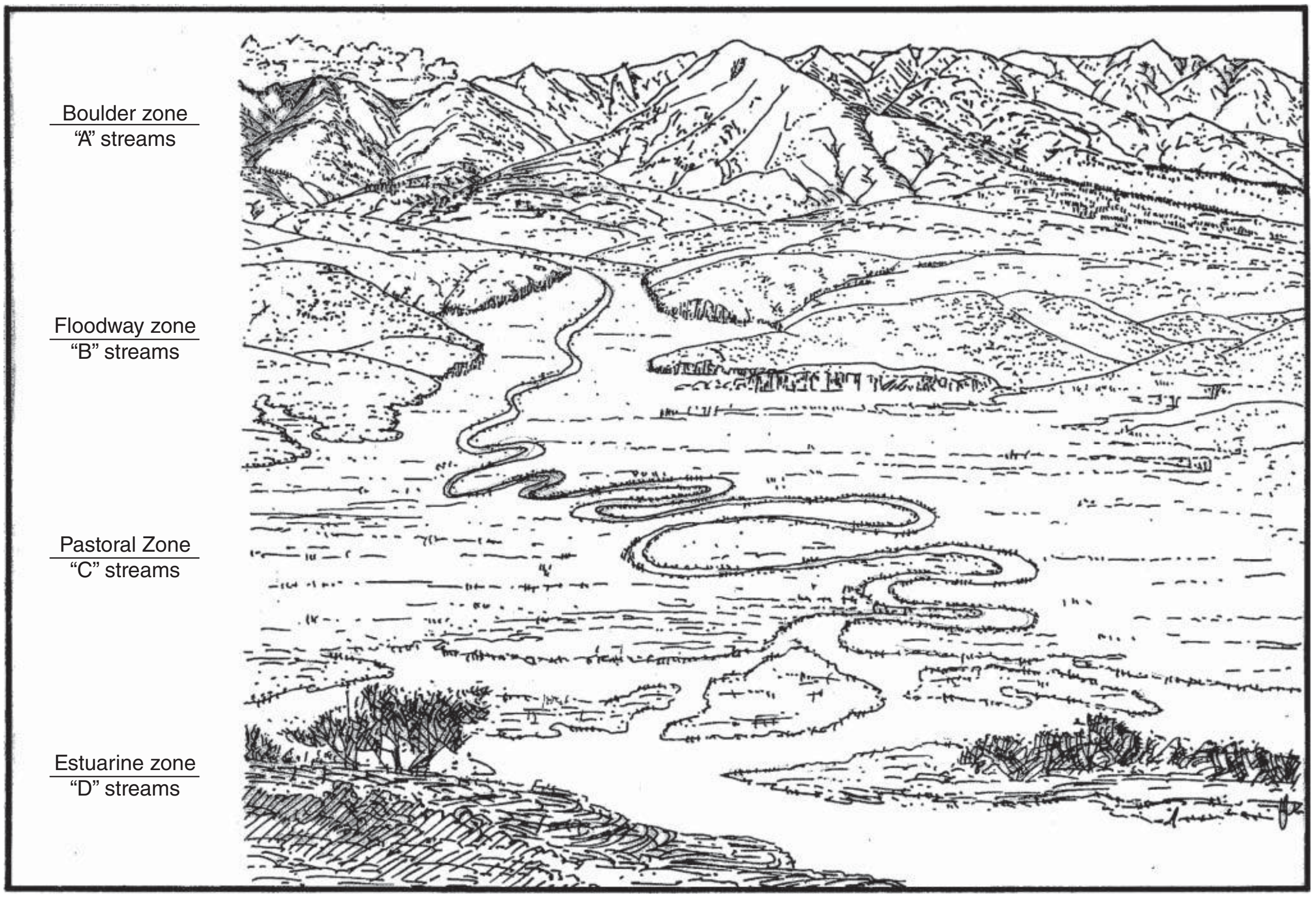


Figure 7.4 Oblique view of idealized river system from headwaters to sea, illustrating the geo-hydraulic zones of Bauer (adapted from Palmer 1976) and corresponding types of Rosgen (1994) and Nevins (1965). Source: Kondolf, 1995. Reproduced with permission of Wiley.

that serve as sources for large boulders can impart substantial local variability—such as forming the rapids along the Colorado River through the Grand Canyon. Similarly, flow obstructions such as logs and log jams can trigger pool scour and bed fining that impart local variability to general downstream changes in channel geometry.

Geomorphic features along a hydrographic network can be gradual or segmented (Leviandier *et al.* 2012). From a geomorphic point of view, depending on the physiographic environment, the boundaries between adjacent stream types can be abrupt or gradual. On a local scale, rivers are also characterized by periodic structures such as pool-riffle sequences and meanders, with characteristic wavelength and amplitude. The longitudinal pattern is therefore a complex structure, usually a hybrid combination of gradual, segmented and periodic signals, usually responding discontinuously at different spatial scales.

Consequently, the appropriate philosophical underpinnings for channel classification inherently depend on the purpose to which it is to be applied, which, in turn, is specific in terms of scale and regional context. Because segmentation is one of the properties of the longitudinal signal, classification of forms can be done without being totally arbitrary.

7.2 Classifications for fluvial understanding

A wide range of geomorphic river classification schemes have been proposed since the late 19th century, reflecting the diversity of environmental settings, the variety of potential approaches to ordering complex natural systems, the intellectual framework of the field and the diverse purposes for which the systems were developed.

Early classifications

Distinctions between mountain torrents and lowland rivers are perhaps the oldest form of river classification. The mineralogist James Dana (1850) offered an elegant description of the difference between mountain streams and lowland alluvial channels based on his experiences scaling the interior of islands in the South Pacific. Powell (1875) proposed a classification of rivers based on their genetic relation to geological structure. At the close of the 19th century, the *geographic cycle* of Davis (1899) fitted neatly into the philosophical notions derived from evolutionary theory then in vogue by fitting landscapes and rivers into stages of an evolutionary cycle. In the late 19th and early 20th century, the relationship of channel network form to geological history, lithology and structure was recognized by the pioneering work of Gilbert (1877) and the classification by Zernitz (1932) of channel network forms based on branching angles as the now familiar dendritic, trellis and radial channel network forms. Stream order (Horton 1945; Strahler 1957) was also a pioneering classification of rivers, providing a longitudinal segmentation based on a relative size index.

Process-based classification of channel patterns

Leopold and Wolman (1957) presented a quantitative basis for differentiating straight, meandering and braided channel patterns based on relationships between slope and bankfull discharge. This early process-based classification has been revisited and enlarged (Rust 1978; Ferguson 1987; Schumm 1985; Church 1992; Thorne 1997; Alabyan and Chalov 1998), in particular with the inclusion of additional patterns such as anastomosing (Smith and Smith 1980; Knighton and Nanson 1993; Makaske 2001) and more generally anabranching rivers (Nanson and Knighton 1996). Schumm (1963, 1977) classified alluvial rivers on the basis of whether their beds are stable, eroding or aggrading and further differentiated them through the dominance of suspended load, mixed load or bedload sediment transport (Table 7.2). Church (1992, 2006) summarized channel patterns based on sediment supply, sediment calibre and channel gradient in a useful diagram, based on the concepts of Mollard (1973) and Schumm (1985), and further refined the diagram (Fig. 7.5). Nanson and Croke (1992) took account of the strong dependence between channel and floodplain to propose a detailed genetic floodplain classification.

Upland channels have been classified based on dominant processes (e.g. Bertrand *et al.* 2013a; Buffington and Montgomery 2013), such as the relative influence of hillslope versus fluvial processes (e.g. Grant *et al.* 1990; Whiting and Bradley 1993) and mechanisms of deposition such as step-pool (Curran and Wilcock 2005). Montgomery and Buffington (1997) found that different mountain channel reach morphologies had different relative transport capacity as expressed in terms of stream power or drainage area and reach slope. Although such distinctions provide for a natural classification of channel types, they arguably represent stratification of a continuum of natural channel morphologies.

With recognition of stream power as a key variable in fluvial geomorphology an increasing number of classifications have been based on this parameter (Table 7.3) (Schmitt *et al.* 2001). Stream power-based classifications have been applied at finer spatial resolutions than the common resolution of channel patterns (Newson *et al.* 1998; Schmitt 2001; Brierley and Fryirs 2005). When different stream power-based classifications are compared, overlaps of the specific stream power classes are frequently observed (Table 7.3). This imprecision can be due to estimations of basic parameters (Schmitt *et al.* 2001), the lack of clear stream power thresholds between channel patterns (Ferguson 1987) due to the influence of other controlling factors (e.g. bedload supply, bank resistance) and the effect of the geographic setting. In most cases, stream power-based classifications are supplemented by geomorphic variables at the levels of valley bottom, floodplain or channel. Moreover, specific stream power (stream power per unit channel width) is not an independent variable, as channel slope depends in part on sinuosity and width depends on channel geometry, which are two dependent variables (Van den Berg 1995). Nonetheless, specific stream power appears to be a useful variable for constructing geomorphic classifications at different spatial resolutions and has the potential to take into account channel processes and adjustments (NRA 1992; Kondolf 1995; Newson *et al.* 1998; Brierley and Fryirs 2005).

Table 7.2 Classification of alluvials based on Schumm (1963, 1977). Reproduced with the permission of Wiley.

Mode of sediment transport and type of channel	Channel sediment (*M*) (%)	Bedload (percentage of total load)	Channel stability		
			Stable (graded stream)	Depositing (excess load)	Eroding (deficiency of load)
Suspended load	>20	<3	Stable suspended-load channel. Width/depth ratio <10; sinuosity usually >2.0; gradient, relatively gentle	Depositing suspended load channel. Major deposition on banks causes narrowing of channel; initial streambed deposition minor	Eroding suspended-load channel. Streambed erosion predominant; initial channel widening minor
Mixed load	5–20	3–11	Stable mixed-load channel. Width/depth ratio >10, <40; sinuosity usually <2.0, >1.3; gradient, moderate	Depositing mixed-load channel. Initial major deposition on banks followed by streambed deposition	Eroding mixed-load channel. Initial streambed erosion followed by channel widening
Bed load	<5	>11	Stable bed-load channel. Width/depth ratio >40; sinuosity usually <1.3; gradient, relatively steep	Depositing bed-load channel. Streambed deposition and island formation	Eroding bed-load channel. Little streambed erosion; channel widening predominant

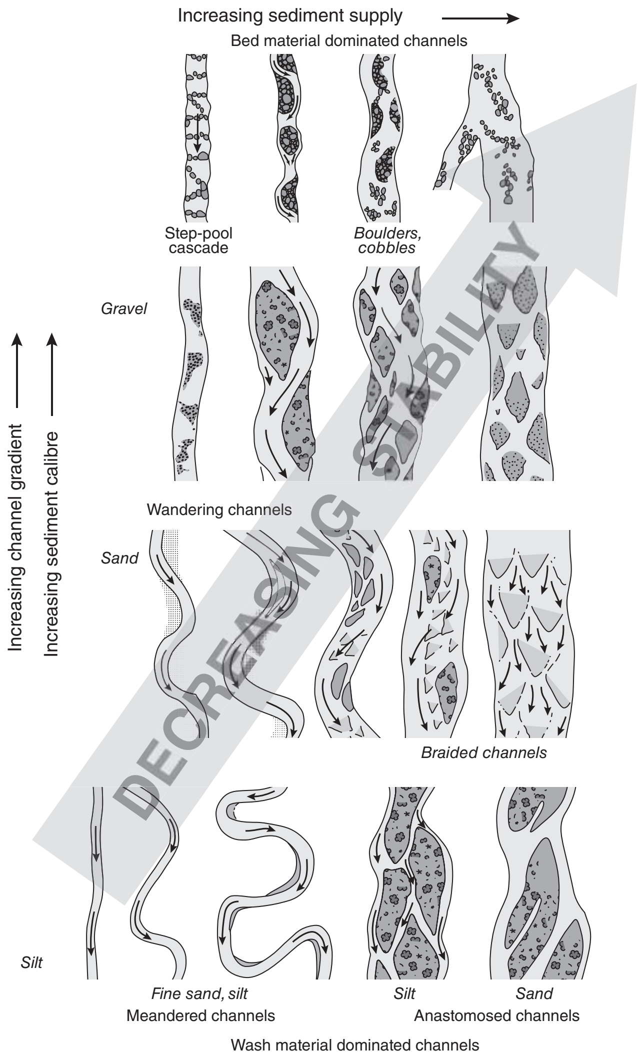


Figure 7.5 Alluvial channel form and its principal governing factors, with shading indicative of sediment character. Copyright reserved by Michael Church, used with permission.

Church (2006) linked alluvial channel morphology to sediment transport and deposition, developing a comprehensive alluvial channel classification based on Shield's number, clearly articulating channel features and forms as reflective of transport processes. Among key distinctions are between threshold channels in which sediment transport is mediated by a surface layer of gravel, versus labile (typically sand-bed) channels, in which bed sediments are frequently and easily mobilized (Church 2006).

Hierarchical classifications

Classification models as determined above lead generally to the definition of *interlocked spatial units* within which the variability of each smaller hierarchical level is constrained by that of the higher hierarchical level (see Chapter 5). At the broadest scale, differences in styles of precipitation and vegetation lead to differences in river processes and characteristics in major climate zones (e.g. alpine, tropical, temperate, arid and polar regions). Ecoregions defined by areas of similar climate, vegetation, lithology and topography (Omernik 1987) can be related to the characteristics of aquatic habitats (Rohm *et al.* 1987; Imhof *et al.* 1996; Allan and Johnson 1997). Just as there are a number of ways to broadly stratify general environmental influences on river systems, there are many ways to address channel classification at finer spatial scales.

In the Loire River basin (100,000 km^2), characteristics of river corridors depend largely on the morpho-region (>100 km^2) drained (Fig. 7.6). For example, in the highlands of the Massif Central (Upper Massif Central and Granitic Plateau of Massif Central), river corridors consist of steep, narrow valleys, whereas in the granitic Armoricain region (Armoricain Massif), they consist of wide, gently sloped valleys (Fig. 7.7) (Bethemont *et al.* 1996). Each morpho-region has a characteristic longitudinal distribution of valley morphology, channel morphology and in-channel features (Souchon *et al.* 2000). For example, in the granitic Armoricain region, second-order streams are mainly characterized by shallow water with moderate velocity and third- to fifth-order streams are dominated by deep waters with low velocity. In the 'sedimentary' region, as streams increase from second to fifth order, the slope decreases and the geomorphological features change from pool-riffle sequences or plane beds to homogeneous, deep, low-velocity channels. This classification yields classes within which channel variability is relatively consistent in each morpho-region of the Loire and provides a tool to manage aquatic ecosystems at the river-basin scale.

Montgomery and Buffington (1997, 1998) proposed a hierarchical valley segment and reach-level classification of mountain channel networks based on morphological attributes related to the ratio of sediment supply to transport capacity, recognizing colluvial, alluvial and bedrock valley segment types. Colluvial valleys are headwater valley segments with relatively ineffective fluvial sediment transport and in which colluvium delivered from hillslopes accumulates as colluvial valley fills. Bedrock valley segments are those in which little material is stored in the valley bottom, whereas alluvial valley segments are those with thick alluvial valley fills. Montgomery and Buffington (1997, 1998) also recognized eight distinct channel reach types that can be used to characterize a continuum of natural channel reach morphologies (Fig. 7.8). These reach types are defined by discrete bed morphologies interpreted generally to reflect relative transport capacity over shorter time-scales than the valley morphologies described above. These channel types are intended to allow comparison of comparable reaches, and although these reach-level channel types are generally correlated with reach slopes, they also reflect local conditions and disturbance history.

The concepts of process domains and lithotopo units implicit in this approach provide for classification at spatial scales greater than individual channel reaches (Montgomery 1999). Process domains are areas of a watershed that are dominated

Table 7.3 Synthetic and comparative representation of some specific stream power-based river classifications. The correlation between the specific stream power classes is rough.

Specific stream power (W m^{-2})	Ferguson (1981, 1987)	Nanson and Croke (1992) (floodplains) (classification 1st level)	Petit (1995)	Nanson and Knighton (1996) (anabranches)	Bernot and Creuzé des Châtelliers (1998) (typology 2nd level)
(+)					+100 < ω < +1000 V-shaped valley confined
		>300 non-cohesive floodplains, high energy			+100 < ω < +1000 U shaped valley deep
	120–300 active low-sinuosity channels		>100 often pattern modifications (braiding is possible)	100–300 Type 6	50 < ω < 500 V-shaped valley widen
				30–100 Type 5	30 < ω < 300 presence of a floodplain
	20–350 confined meandering	10–300 non-cohesive floodplain medium energy		50 to 5–10 Type 3	30 < ω < 700 channel limited by incision
	5–350 active meandering		<35 no self-adjustment after regulation	15–35 Type 4	30 < ω < 120 large floodplain
(–)	1–60 inactive channels	<10 cohesive floodplains low energy	<15 inactive channels	≤8 Type 1 4–8 Type 2	ω < 30 littoral floodplain

Explanation of Nanson and Knighton's anabranch types:

Type 1: cohesive sediment anabranching rivers.
Type 2: sand-dominated, island-forming anabranching rivers.
Type 3: mixed-load, laterally active anabranching rivers.
Type 4: sand-dominated, ridge-forming anabranching rivers.
Type 5: gravel-dominated, laterally active anabranching rivers.
Type 6: gravel-dominated, stable anabranching rivers.

Source: Schmitt *et al*, 2001. Reproduced with permission of Zeitschrift für Geomorphologie

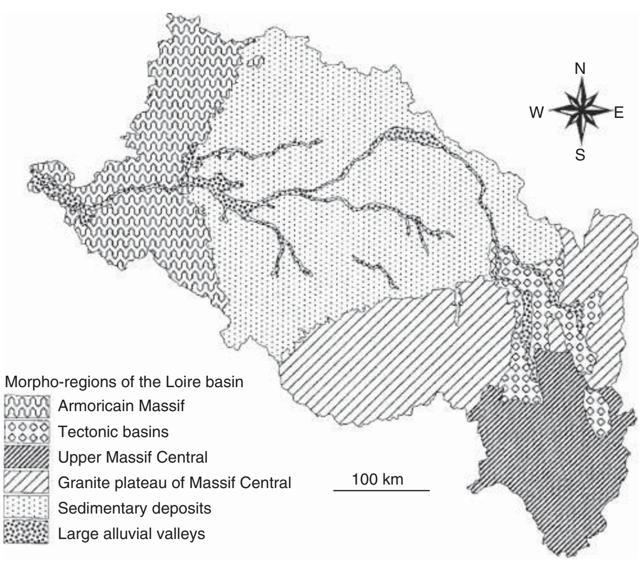


Figure 7.6 Distinct morpho-regions of the Loire River basin identified by Bethemont *et al.* (1996). Source: Bethemont *et al*, 1996. Reproduced with permission of Revue de Geographie de Lyon.

by comparable geomorphological processes and therefore with similar sediment transport dynamics and comparable disturbance regimes. Channels within a process domain would be expected to experience similar disturbance processes and different process domains roughly delineate a longitudinal channel classification (Fig. 7.9).

Integrating temporal trajectories in classification schemes

Alluvial channels are continuously adjusting over decades, centuries and millennia to the evolution of independent variables (i.e. flow and sediment supply), which vary continuously due to variable climate, natural and anthropogenic landcover change, neotectonic activity and engineering works (Brierley and Fryirs 2005). Moreover, in many areas, present channel dynamics are controlled, at the spatial scale of the floodplain, by inherited geomorphic features such as glacial and fluvio-glacial deposits, legacy of fluvial palaeodynamics. These can, respectively, limit lateral dynamics due to the presence of boulders exceeding the river competence or favour channel narrowing when a channel

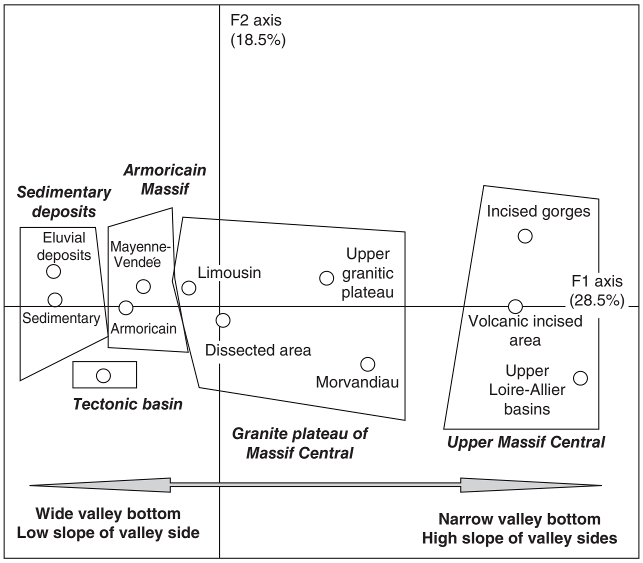


Figure 7.7 Distribution of the 12 elementary morpho-regions of the Loire River basin on the first factorial map F1–F2 of a multiple correspondence analysis using five morphological variables (stream order, valley slope, valley side slopes, channel sinuosity, low flow channel width) by Bethemont *et al.* (1996) in the Loire River basin. The regions line up along the F1 axis in five main regions, from the sedimentary region near Nantes with wide alluvial valleys and gentle side slopes to the incised volcanic region of the Massif Central with narrow valleys with steep side slopes. See Fig. 7.6 for location of the regions. Source: Bethemont *et al*, 1996. Reproduced with permission of Revue de Geographie de Lyon.

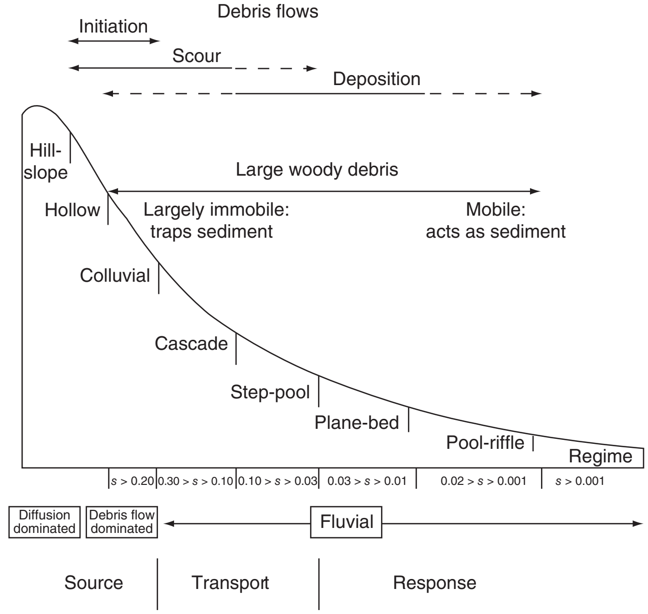


Figure 7.8 Distinct channel reach types of Montgomery and Buffington (1997, 1998), shown as a function of transport or supply limitation. Source: Montgomery and Buffington, 1997. Reproduced with permission of Geological Society of America.

abandoned by a big river after an avulsion is fed only by groundwater or reoccupied by a smaller tributary, as is the case for some channels on the alluvial plain of the Rhine (Schmitt 2001; Schmitt *et al.* 2007) (see Fig. 7.12). Therefore, alluvial plains can be viewed as palimpsests, with fluvial legacies increasing the complexity of channel–floodplain interactions (Bravard and Gilvear 1993).

Integrating time in functional classification schemes is possible through the types of channel adjustments as proposed by Brookes (1987) and Downs (1995), by studying old maps (Kondolf and Larson 1995) or by taking into account sediment supply evolution in comparison with the present transport capacity (Montgomery and Buffington 1998). However, in many cases it is essential to consider channel dynamics in longer temporal trajectories covering several centuries or millennia, in some cases since the beginning of the Holocene, in order to evaluate more accurately recovery potential and sustainability of present and future management strategies, including restoration and ecology (Bravard *et al.* 1986). The increasing availability of data concerning fluvial palaeodynamics and alluvial archaeology is an important help in this respect (Brown 2002).

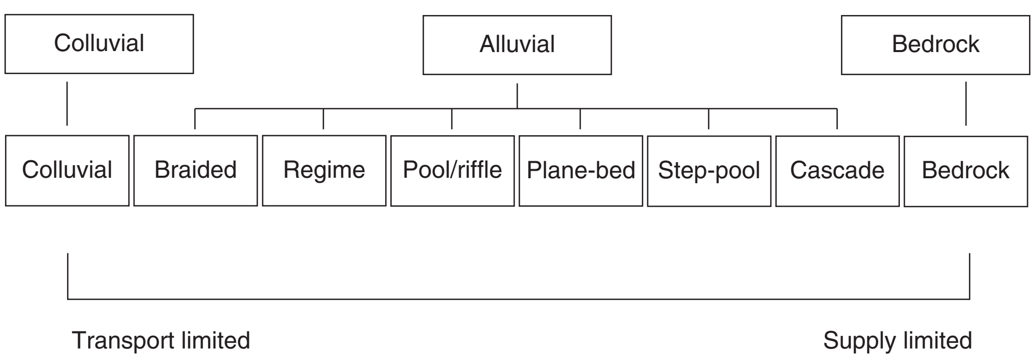


Figure 7.9 Process domains of Montgomery and Buffington (1997) arranged along a longitudinal gradient. Source: Montgomery and Buffington, 1997. Reproduced with permission of Geological Society of America.

7.3 Interactions between geomorphic classifications and ecology

As fluvial geomorphology (forms and processes) is a key component of aquatic physical habitats, it is clear that it exerts a strong control on aquatic ecological communities and processes (Hynes 1975; Newson and Newson 2000). Links between fluvial geomorphology and ecology have been displayed in many studies at different scales, for both fauna (Huet 1949; Hynes 1975; Beisel *et al.* 1998) and flora (Carbiener 1983; Holmes 1989). Similarities between the longitudinal dimension of the geomorphic concept of Fluvial System (Schumm 1977) and the ecological concept of River Continuum Concept (Vannote *et al.* 1980) have been highlighted by several authors (e.g. Amoros and Petts 1993). These similarities concern energy, water and particle fluxes, vegetation (terrestrial and aquatic), invertebrates and fish. Cupp (1989), Bostelmann *et al.* (1998b) and Chessman *et al.* (2006) demonstrated significant differences between biological patterns and geomorphic stream types. Predictive models for invertebrate communities notably based on geomorphic data have also been developed (Wright *et al.* 1984; Ferréol *et al.* 2008).

Riquier *et al.* (2015) identified distinct backwater channels according to overbank flow frequency and shear stress conditions and showed that types are ordered along a connectivity gradient, which ecologically structures macroinvertebrate communities (Castella *et al.* 2015). Belletti *et al.* (2013) identified six braided channel types, distinguishing P pond, AL alluvial channels, SC secondary channels, MIX upstream connected channels, MC main channel, GW groundwater channels, and calculating Shannon diversity indices (H') for each based on their relative lengths. Wawrzyniak *et al.* (2013) demonstrated a relationship between H' and the thermal range in a braided channel network. Following these findings, H' and discharge frequency provided a basis for identifying the most valuable reaches in term of aquatic habitat (i.e. with the highest H', type 2) and their locations with the southeastern part of the French Alps (Belletti *et al.* 2013) (Fig. 7.10).

Other recent work has focused on interactions between hydromorphic process-based typologies and ecological characters and functioning. Schmitt *et al.* (2011) demonstrated in different hydrosystems relationships between a functional geomorphic typology (Schmitt *et al.* 2004, 2007) and aquatic macrophyte communities in the French alluvial plain, and also benthic and hyporheic oligochaete assemblages in the Yzeron River, a peri-urban catchment impacted by combined sewer overflows (CSOs) near the city of Lyon (France). In both settings, physical and ecological relationships are controlled by surface

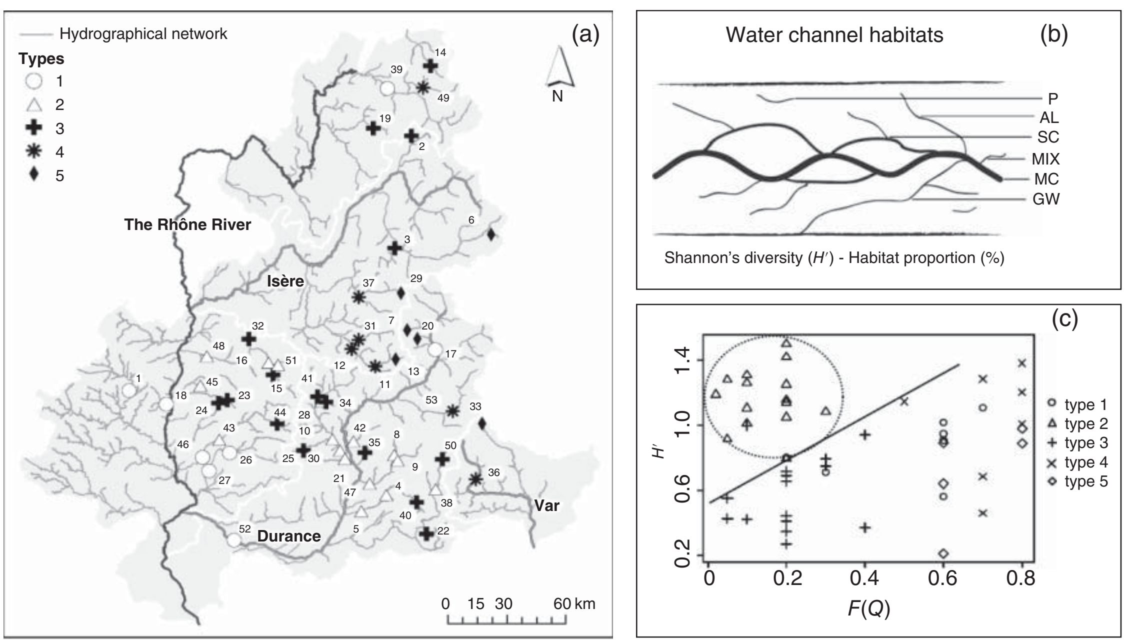


Figure 7.10 Relations between process-based features and ecological characteristics of the Alpine French braided rivers. (a) Typology of reaches based on longitudinal gradient, elevation, slope, sediment availability and active channel width. Types 1, 2 and 3 are lowland braided reaches. Types 1 and 3 both have narrow active channel widths, but differ in how recently they experienced scouring high floods, with type 1 reflecting recent high flow. Type 2 reaches have wider active channels than types 1 and 3 due to higher sediment supply. Types 4 and 5 are situated mainly in the inner Alps and have higher summer flows due to ice and snowmelt. Type 4 reaches occur downstream of type 5, with lower elevation and gradient. (b) Aquatic habitats in braided channels: P = pond, AL = alluvial channels, SC = secondary channels, MIX = upstream connected channels, MC = main channel, GW = groundwater channels. (c) plot of H' (Shannon diversity index of channel habitats based on their cumulative length) versus exceedance flow frequency, $F(Q)$, for different braided channel types shown in (a), showing higher values of H' for braided channel type 2. The sloping line distinguished groundwater-fed type 2 from the other types. Source: Belleti *et al*, 2013. Reproduced with permission of Springer.

water–hyporheic water exchanges. In the most active channel type of the Rhine, upwelling of nutrient-poor groundwater reduces the growth of algae that might otherwise clog channels and results in longitudinal changes in macrophyte communities and increased diversity of communities. In the Yzeron basin, the severity of impacts of CSOs is largely mediated by processes determined by geomorphic typology. Geomorphic types with higher values of energy and grain size favour nutrient cycling and self-purification processes, by virtue of greater water turbulence and dynamic surface water–hyporheic water exchanges (Schmitt *et al.* 2011). These examples demonstrate that it can be possible to predict spatially ecological characteristics and functions on large spatial scales through functional geomorphic classifications.

7.4 Geomorphic classification and quality of river environments

Management-oriented classifications are often used to assess environmental quality (or state of degradation) of reaches as an aid to prioritizing for intervention. Increasingly, river management is undertaken at the catchment scale or at least over river reaches several kilometres in length, with recognition of upstream–downstream interdependence to provide insights into the response potential of some river systems, such as identifying reaches that have been historically unstable (Brookes 1987; Downs 1994) or prone to lateral channel migration ('erodible corridors') (Piégay *et al.* 2005) and reaches more sensitive to effects of upstream land-use changes (Downs 1995; Montgomery and Buffington 1998), as for example channel adjustment downstream of urban inflows and combined sewer overflows (Grosprêtre 2011). Based on the finding of Brookes (1987, 1990) that habitat enhancement structures in Denmark and Britain typically did not survive in channels subject to a unit stream power exceeding $35\,W\,m^{-2}$, the National Rivers Authority (now the Environment Agency) included an estimate of stream power in its classification system as a guide to the likely impact of channel modification (NRA 1992). Classification can provide a preliminary indication about whether various bank protection works and habitat enhancement structures are likely to prove successful. Such schemes can also identify homogeneous spatial patterns for river management purposes, such as scenarios based on sediment reintroduction (Bertrand *et al.* 2013b), or to prioritize river restoration based on a mapping of historical changes as done for the Bega River, Australia (Brierley and Fryirs 2000).

Classification can be a useful tool in stream restoration in at least three ways: (1) surveying existing conditions and setting priorities for restoration, (2) envisioning an end state towards which restoration should proceed and (3) providing initial indications about restoration measures likely to succeed in a given channel (Kondolf 1995). Classification has been heavily used as a basis for restoration, despite the limited scientific basis of some systems and the disappointing performance of many of the resulting projects.

The adoption of the EU Water Framework Directive in October 2000 (European Parliament and Council of Europe 2000) has significantly promoted the development of typologies of river reference status, that account for alteration by human activities. This Directive aimed to achieve 'good ecological status' for surface and groundwater resources of all member states by 2015. The approach is to implement detailed action programmes of regulatory and economic measures within hydrographic basins. The Directive is profoundly modifying national water policies in the member states by setting up methods for measuring and characterizing each category of surface water body: rivers, lakes and transitional and coastal waters (Raven *et al.* 2002; Piégay *et al.* 2008). For rivers, member states must make comprehensive maps of chemical and ecological status, including hydromorphological status as reflected in flow regime, sediment transport, river morphology and lateral channel mobility. Member states are required to monitor river ecological quality (biological, hydromorphological and physicochemical quality) and, if necessary, develop restoration programmes.

Reviewing about 140 assessment approaches to characterize biophysical river features worldwide (Table 7.4), Rinaldi *et al.* (2013) identified five broad categories of hydromorphological assessment methods: physical habitat, riparian habitat, morphological, hydrological regime alteration, and longitudinal fish continuity. Half of the methods focused on the physical habitats, with only 16% including a morphological assessment. Most of the habitat-focused methods emerged in the 1980s–1990s (e.g. Platts *et al.* 1983; Raven *et al.* 1997). These management-oriented assessment approaches were mainly single-scaled, focusing on channel reaches along a river continuum or sampled randomly in a given area.

These approaches raise important questions about 'reference conditions' (Dufour and Piégay 2009; Morandi *et al.* 2014). Classification can help to stratify channels that should have similar reference conditions against which to assess the degree of degradation and provide an indication of the historical (pre-disturbance) condition at the site to inform restoration designs. The historical condition is not necessarily a suitable goal for restoration design, as any such ideal must be adjusted to account for irreversible changes in controlling factors (such as runoff regime and sediment supply) and for considerations based on cultural ecology at the site, such as preservation of historical land uses or creation of habitat for endangered species. From this discourse emerges the understanding that there may be different kinds of references: historical; geographical; giving a value to naturalness (as opposed to wilderness) and using the most natural reaches in a given region as references; or functional, using as reference river reaches that are functioning well (e.g. no real constraints in terms of bedload transport, erosion or flooding, so that forms are self-sustaining). This functional approach to 'reference' fits well within the concept of

Table 7.4 Censing of existing hydromorphological assessment methods according to five broad categories and countries.

Location	Physical habitat	Riparian habitat	Morphological assessment	Hydrological assessment	Fish continuity	Total
Europe	40	5	12	4	13	74
United States	24	5	8	4	5	46
Austria	6				1	7
Belgium	2				2	4
Czech Republic	1		1			2
Denmark	5					5
England & Wales	4		4		2	10
France	3		2		2	7
Germany	5				1	6
Ireland (NI and RoI)	1		1			2
Italy	2	1	1	1	1	6
The Netherlands	1				1	2
Poland	3		1			4
Portugal	1					1
Scotland			1	1	1	3
Slovakia	1					1
Slovenia	1					1
Spain	2	4	3	2	2	13
Sweden	2					2
Switzerland	1					1
Australia	4	2	1			7
Others*	4	2	2	2	2	12

*South Africa, Canada/Quebec, China, New Zealand, Taiwan, Ukraine.
Modified from Rinaldi *et al.* (2013).

'rehabilitation' as opposed to 'restoration' as explained by Henry and Amoros (1995).

Channel classification can provide a 'guiding image' or *Leitbild*, of the channel form that would naturally occur on the site, adjusted to account for irreversible changes in controlling factors (such as runoff regime) and for considerations based on cultural ecology (such as preservation of historical land uses or creation of habitat for endangered species), a concept pioneered in Germany (Kern 1992, 1994). Attributes of this ideal channel form can be adopted as goals for restoration projects. Thus, the *Leitbild* is a model of the ideal channel design for a site based on physical and ecological considerations, including historical changes to runoff and sediment yield. Based on constraints such as flood control, pre-existing water rights and budget limitations, planners propose an optimal design for review by resource agencies and the public and which is ultimately modified into a feasible design for the site (Kern 1992).

The *Leitbild* concept has been applied extensively in Germany as a basis for assessing existing channel conditions and to provide guidance for restoration, for regions with similar geology, climate, etc., in which a consistent set of valley and channel forms could be expected. The characteristics of various *Leitbild*s have been defined through detailed field study and practitioners are encouraged to visit illustrative reaches displaying properties of the *Leitbild* to aid the development of more compelling conceptual models for restoration efforts than can be gleaned from diagrams and statistics alone. The *Leitbild* approach requires some judgement (and thus professional background) to apply, largely because historical changes in basin and channel conditions must be understood and considered in developing the *Leitbild* for a given site. The process is more than simple mimicry of remnant natural channels found in undisturbed drainage basins. Such channels provide an indication of the potential natural state of a given class of channel and attributes that might be considered for restoration objectives, but cannot indicate how to address constraints such as altered runoff patterns when developing a *Leitbild* for the project reach. Similarly, historical reconstructions of former conditions in the subject reach provide useful insights into the ecological potential of the site, but historical conditions may be impossible to recreate and maintain because of changes in the catchment.

In addition to the issues with the concept of 'reference', debates have also concerned how to approach or infer processes, with its implications for the potential sensitivity to changes in independent variables. Pioneers in this domain were the ecologists who sought to identify physical factors affecting biological communities at the habitat scale. The most ambitious effort was the River Habitat Survey (RHS) system, which was developed in 1994 to provide a unifying basis for river classification and evaluation in the UK (Raven *et al.* 1998). RHS comprised four related outputs (Raven *et al.* 1997): (1) a standard field survey method, (2) a large computer database, (3) a classification of unmodified rivers based on a physical predictive model and (4) a technique for assessing river habitat quality. The RHS data set comprised 17,000 sites in England, Wales, Scotland and Northern Ireland (Raven *et al.* 1997), located on the basis of a random stratified sample. Additional sites located in different European countries have been added more recently (Raven *et al.* 2002). Each site was 500 m long and extended 50 m either side of the channel (Raven *et al.* 1997). The variables measured included the physical structure of rivers, using the habitat level as the basic element for river management (Harper and Everard 1998). The data were obtained from maps and stream gauging records (e.g. altitude, slope, geology, distance from source, mean annual flow) and field surveys (e.g. width, depth, channel substrate and geomorphological units, bank vegetation structure and artificial modifications such as weirs, dams, bank reinforcement, channel deepening or realignment). The data were integrated in a computer database to establish a reference network of relatively undisturbed river sites (Raven *et al.* 1997). A key element of RHS was the elaboration of a semi-natural hydromorphological river typology derived from this subset of reference sites, which support comparison of a site to 'reference' conditions and allow a given site to be assessed in the context of all sites of the same river type (Environment Agency 2002). This assessment comprises a simple notation in five classes (excellent, good, fair, poor, bad), which reflects the deviation from the reference state given by the river typology (Raven *et al.* 1997, 1998).

It is instructive that this classification scheme was developed iteratively. First, a classification of 11 types resulted from statistical analysis, distinguishing sites according to geology, altitude, slope and mean annual discharge (Raven *et al.* 1997; Environment Agency 2002). This scheme was abandoned because intra-type variability equalled or exceeded the variability between the types. A second classification of nine types was elaborated, but it did not adequately predict different habitat features. A third iteration of the classification was based on the observation that most geomorphological features were correlated with map-based variables such as altitude, slope, distance from the source, altitude of the source and geology. A principal component analysis on the table 'individuals–variables' (individuals being the surveyed stations and the variables being the field measures) simplified the initially large set of variables into a smaller set of synthetic variables (principal components). In this case, the two first components (F1 and F2) explained 90% of the total variance (or inertia). The first component represented a gradient of altitude and slope, while the second component was correlated with discharge and reflected a potential 'energy' gradient (Jeffers 1998; Environment Agency 2002). A summary of the different river characteristics can be viewed on the first factorial map, which shows that the limits between groups were not defined clearly (Fig. 7.11). To facilitate understanding, the biplot was divided by lines drawn arbitrarily across a continuum to define eight named 'types'. The semi-natural features could be predicted by the four map-based variables or by scores on the two principal components (Jeffers 1998; Environment Agency 2002). In a complementary approach, using RHS data and other

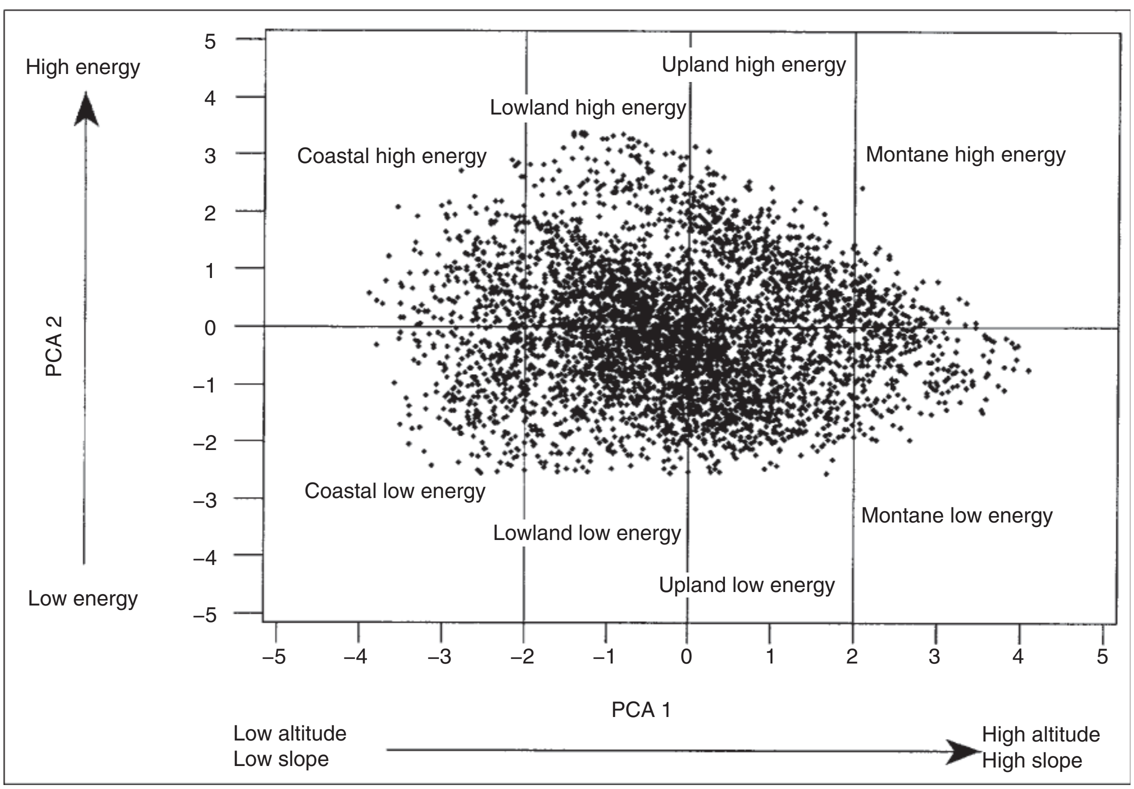


Figure 7.11 Principal component analysis performed on 4569 English and Welsh sites described by their altitude, slope, distance from source and altitude of the source, highlighting the semi-natural river typology of the RHS approach. Source: Environment Agency, 2002.

process-related variables, Newson *et al.* (1998) highlighted the basis of a dynamic river classification and emphasized the importance of specific stream power as a variable for classifying rivers, although this improved the prediction of class type by only 17%. Links with vegetation have also been developed (Erba *et al.* 2006).

Most approaches identified by different member states as their procedures for European WFD-mandated hydromorphological assessment are unfortunately very limited for understanding processes, adjustment patterns and linkages between changes (e.g. alterations) and potential causal factors. These limitations have inspired development of new methods since 2000 that consider more broadly the geomorphic conditions and adjustment conditions, such as the IHG (Indice Hydrogeomorfologico) (Ollero *et al.* 2007, 2011) and the MQI (Morphological Quality Index) (Rinaldi *et al.* 2013). These new indices value naturalness and also historical condition and devalue changes occurring over recent decades, implicitly assuming that river states before adjustment were more valuable. However, after a river has changed, it may not continue to change because it may have already adjusted. The sensitivity of a river to future change is not yet considered. The MQI is based on 28 indicators, characterizing longitudinal and lateral continuity, channel pattern, cross-section configuration, bed structure and substrate and riparian vegetation. The method is process based, taking into account sediment continuity and wood flux, bank erosion, lateral mobility and channel adjustments (Table 7.5). These parameters are analysed in terms of geomorphic functionality, artificiality and channel adjustments. Indicators, classes and the scoring system are based on expert judgement. The procedure is also based on a preliminary typology of rivers (confined, semi-confined, unconfined) so as to adapt the scoring system to the geomorphic context. This geomorphic quality assessment is based on a preliminary phase of river reach segmentation, consisting of an initial division of the network into river reaches with homogeneous geomorphic patterns.

The SYRAH (Système Relationnel d'Audit de l'Hydromorphologie des Cours d'Eau) approach used in France (Chandesris *et al.* 2009) is based on some of these paradigms (e.g. valuing naturalness, integrating longitudinal continuity) focusing on the risk of alteration due to human pressures on the river network in terms of bedload and suspended load transfers, morphological status based on presence of infrastructures such as roads, weirs and bridges and simplification of channel geometry. The number of parameters and scoring system are not as integrated and advanced as that of Rinaldi *et al.* (2013), but the variables are summarized in a set of maps, which themselves can inform decision-making (Fig. 7.12).

Future efforts are needed to explore some key aspects of classification: (i) to anticipate better how sensitive/reactive it is in a given reach and to assess whether we can expect it to change in the coming decades – there is no reason to expect the position of each reach in a classification to be static, which has important implications for management; (ii) to link better the reach characteristics with upstream (and downstream) influences and which have a critical influence on the reach (e.g. a channel feature 100 km upstream or patterns within the entire basin), drawing attention to linkages between pressures and impacts (Rinaldi *et al.* 2013).

Table 7.5 Channel adjustment indicators, classes and scores employed in the MQI (Morphological Quality Index) system of Rinaldi *et al.* (2013). Reproduced with permission of Elsevier.

Indicator	Classes	Score
CA1	A – absence of changes in channel pattern from 1950s	0
	B – change to a similar channel pattern from 1950s (*PC-U*) or change of channel pattern from 1950s (*C*)	3
	C – change to a different channel pattern from 1950s (only *PC-U*)	6
CA2	A – absent or limited changes (≤15%) from 1950s	0
	B – moderate changes (15 ÷ 35%) from 1950s (*PC-U*) or changes >15% from 1950s (*C*)	3
	C – intense changes (>35%) from 1950s (only *PC-U*)	6
CA3	A – negligible bed-level changes (≤0.5 m)	0
	B – limited or moderate bed-level changes (0.5÷ 3 m)	4
	C1– intense bed-level changes (>3 m)	8
	C2 – very intense bed-level changes (> 6 m)	12

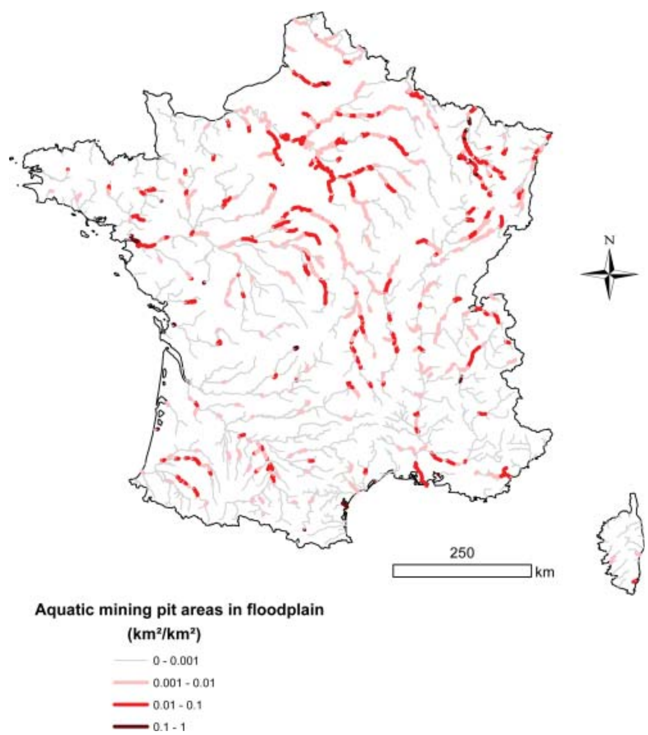


Figure 7.12 Risk of alteration of bedload transport based on the ratio of mining pit areas to floodplain areas based on the database BD Carthage®. Source: Chandesris *et al*. 2009. Reproduced with permission of French National Institute of Geography.

7.5 Applying geomorphic classification schemes to fluvial systems

Before applying classification schemes, several issues must be addressed. Foremost is whether to use an existing classification (and, if so, which one) or to develop a new system for the region and/or problem at hand. This is not a trivial question, as the selection of variables as the basis for a classification implicitly assigns greater importance or significance to the variables used than to the variables not used. The selection of variables is also influenced by the degree to which they lend themselves to measurement and some geomorphically significant variables may not lend themselves readily to quantification.

Procedures for geomorphological data collection in stream channels have been developed independently of stream classification programmes (e.g. Mosley 1982; Hicks and Mason 1991), as part of classification programmes (Kellerhals *et al.* 1976; Jowett and Duncan 1990; Downs and Brookes 1994; Thorne 1998) and incorporated into some ecological data collection protocols (e.g. Platts *et al.* 1983, 1987; Biggs *et al.* 1990). In general, stream inventory procedures include variables such as channel pattern and sinuosity, channel dimensions, bed material size, channel gradient, whether the channel is alluvial or bedrock-controlled, degree of entrenchment and catchment variables such as drainage area, basin relief, valley gradient, annual rainfall and lithology of the basin. These variables, along with indices of channel entrenchment and valley confinement, are readily measured in the field and some can be measured from maps and aerial photographs. If an existing classification is to be applied, then pre-existing classes can be used. If a new classification is being developed, then the number of classes may be determined by the scale of the river system, considerations such as the need to divide responsibility for rivers among different levels of government or the desire to keep the number of classes small enough that the classification system is practical and useful.

In any system designed for broad application by users besides the scheme's author, the selection of variables will be influenced by the availability of data or at least the degree to which certain variables lend themselves to measurement and quantification. For example, bedrock lithology underlying the catchment and the reach can exert profound control on river form, not only through direct effects in bedrock-controlled reaches, but also by controlling the valley walls bounding alluvial reaches, influencing the runoff and sediment load delivered to these reaches and influencing groundwater–surface water interactions (Montgomery 1999). However, there is no simple way to incorporate lithology into most channel classification schemes. Unlike channel width, for example, there is no single number that can represent the range of physical attributes associated with different rock types and structure, such as hardness, permeability, stratification, foliation and fracturing. More feasibly, one can use underlying rock type as one basis for defining homogeneous regions for which a classification scheme can be developed (Kern 1992; Bethemont *et al.* 1996) or within which one would expect similar channel types or longitudinal sequences of types, to exhibit similar finer-scale characteristics (Montgomery 1999). However. even with expert schemes, owing to the great diversity of river dynamics from one region to another, it is possible to learn new things about the behaviour of the rivers in question (and thereby revise the classification scheme) if the scheme's performance is objectively analysed.

Data collection as distinct from identifying channel type

Data collecting and recording should be distinct steps from identifying channels as belonging to a particular class. If not, the observer's perceptions may be influenced by expectations that a channel will fit into a particular class. Essentially the same problem in the context of correlating river or marine terrace remnants based on elevation was discussed by Johnson (1944), who noted a tendency for workers to reach premature conclusions about the suite of terraces present in a given area and rounding elevations of subsequently observed terraces to the nearest preconceived terrace elevation. This resulted in spurious correlations and loss of the real data potentially available if actual values of terrace remnant elevation had been recorded.

> A study on the Atlantic coast reports terraces at 100-foot intervals … [and a subsequent] French investigator has reported successive terraces at intervals of 100 meters. We must conclude either that a wise Providence not only pays remarkable attention to the magic number 100, but also nicely adjusts uplifts of the land and lowering of sea level to the particular system of measure prevailing in each country; or else that actual elevations of very different character are by observers so roughly approximated to the nearest even figure as to make them valueless for correlation purposes. The latter interpretation seems the more reasonable.
>
> *(Johnson 1944, p. 806)*

In using any channel classification, the performance of the classification system should be assessed once the streams have been identified as belonging to specific classes. How many channels were not accommodated by the pre-existing categories? Does the taxonomy need to be modified to accommodate local/regional conditions or the needs/purposes of the project? In other words, channel classification should be used as a flexible tool to help organize understanding, but it should not be blindly relied upon to reach conclusions that may or may not be appropriate to the specific local or watershed context of the channels in question.

Tools used to classify spatial units from data

Most geomorphic classification systems originated as expert systems based on general principles and experience with rivers in a given region. Such classifications 'provide a weak form of explanation because all schemes involve a set of criteria which relate to an *a priori* expectation of the way in which researchers believe their river channels to be distinguished' (Downs 1995, p. 348). Such *a priori* classification schemes reflect the training and experience of the scheme's author, both in the selection of variables to include and classes proposed. An alternative approach is to collect large data sets and employ statistical methods (e.g. cluster analysis) to define objectively patterns or groupings of similar spatial features. In the latter approach, expert judgement is still involved in selecting the variables to use as the basis for the classification scheme, but identification of distinct groupings is left to an objective procedure, although the boundaries may still be subjectively drawn.

Because large sets of sites and variables to describe them are needed at a regional or nation-wide scale to develop a classification using statistical tools, multivariate methods are becoming popular. It is often efficient to summarize the data set variability into main factors (i.e. components of a principal component analysis) and, then perform the cluster analysis (e.g. hierarchical ascendant classification, *k*-means algorithm or others techniques) on these factors (Hallot 2010; see Chapter 21). Newson *et al.* (1998) applied a twin-span analysis to geomorphic and ecological data from 432 sites around the United Kingdom and found no 'objective taxomony', but noted that the river types indicated were intuitively realistic and might prove to be statistically definable with a larger data set. In New Zealand, Mosley (1982, 1987) conducted a cluster analysis on geomorphological data from 190 river reaches and found only four clear interpretable clusters. Nor were clusters evident in a three-dimensional plot of width/depth ratio, channel slope and mean bed material size from 100 rivers measured by Jowett and Duncan (1990). Mosley concluded that a multivariate approach to characterizing rivers was more useful than a classification system for predicting ecological communities supported by the stream and for predicting environmental impacts. Although it may not be possible to identify discrete river types in data sets drawn from many landscape provinces, the approach is likely to be more successful when applied within a given physiographic unit, where multivariate approaches can provide results with practical value for management, as done on Upper Rhine floodplain anabranches (Fig. 7.13) (Schmitt 2001). The dendrogram separates clearly 'anabranches with moderate or no dynamism' from 'dynamic anabranches'. The first group corresponds exclusively to groundwater-dominated channels being palaeo-channels of the Rhine or the Ill River, which do not experience floods and in which morphodynamic activity is weak (narrowing in some cases started several millennia ago). The channels in the second group, which are generally not palaeo-channels, receive floodwaters from the Ill River and thus have morphodynamic activity.

Emergence of data mining: the end and beginning of classification?

River characterization and process understanding are entering a new era with the increased availability and quality of aerial imagery, such as the Pleiades satellite imagery now available 5–10 times per year at a resolution approaching that of traditional orthophotographs (0.5–1 m) formerly acquired every

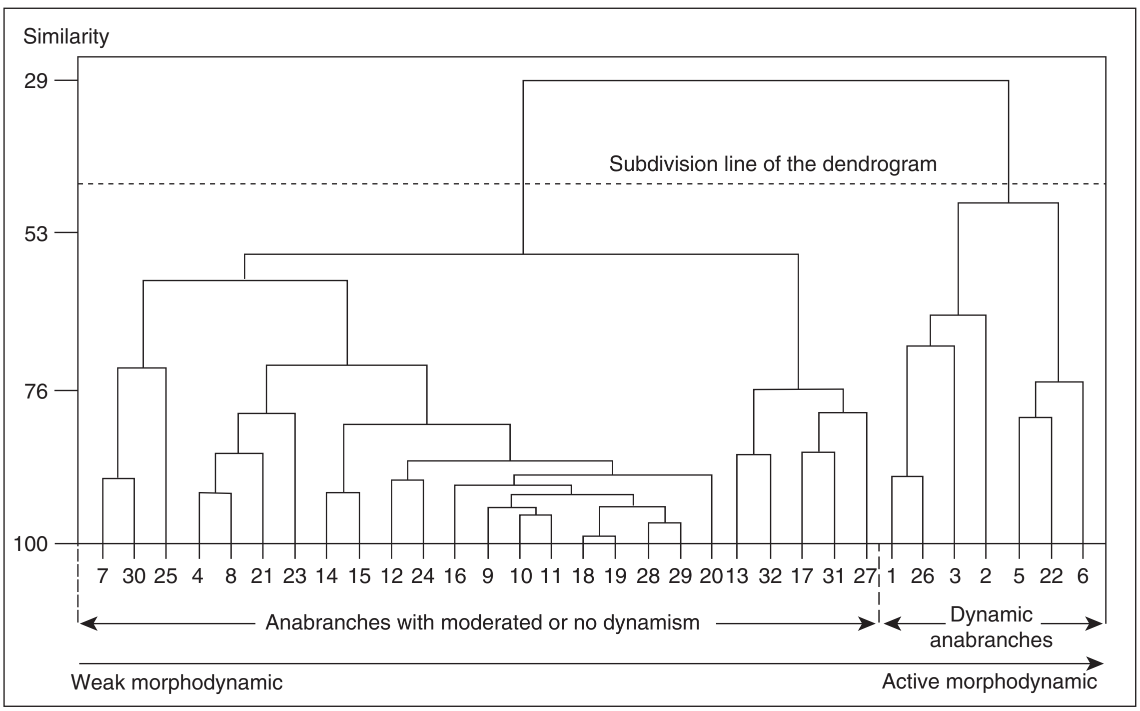


Figure 7.13 Classification of anabranching rivers in the French Rhine River floodplain based on a cluster analysis of 32 sites (numbers). Seven hydro-geomorphic variables are taken into account: sinuosity, width/depth ratio, mean and variation coefficient of in-channel coarse sediments *b*-axis, percentage of lotic morphodynamic units and two lateral mobility indices. The length of a reach corresponds to 36 bankfull width. Source: Schmitt *et al*, 2001b. Reproduced with permission of GéoCarrefour.

3–10 years in many countries. Moreover, with LiDAR surveys, not only the planimetric but also elevation data now permit the calculation of many geomorphic variables.

Remote sensing techniques are therefore becoming critical in fluvial geomorphology (Piégay *et al.* 2012). An ArcGIS River Corridor 'toolbox' provides technical and conceptual methods to acquire and analyse remotely sensed geomorphic data, allowing the user to develop classifications tailored to specific objectives and providing calculation and storage capacities to automate collection and analysis of remotely sensed data (Roux *et al.* 2015). Classifying and delineating features from these data calls for new statistical approaches, such as the segmentation algorithms of Leviandier *et al.* (2012), such as the Hidden Markov Model and Hubert tests (see Chapter 21). In a sense, classification is a working tool to highlight spatial organization and explore theoretically how scale levels are nested within one another. It opens up new possibilities in riverscape analysis to predict how channels can evolve based on a Location for Time Substitution approach (Chapter 5), nicely exemplified by Schumm's (1985) model, expanded at network scale by Brierley and Fryirs (2005) in Australia and by Pont *et al.* (2009) and Bertrand *et al.* (2013b) on the Drôme River, France. The main aim is now to produce data to enrich the regional database to develop links between observed forms and controlling variables, both local and upstream. The integration of field data such as grain size remains for the moment a limitation and a challenge for the future (Buffington and Montgomery 2013).

Limitations and misuse of classification in fluvial geomorphology

Many classification schemes have recently been commissioned or adopted by river management agencies with the aim of simplifying geomorphological analysis to assist in management. While geomorphic stream classification has been around for a long time and (as documented in this chapter) a wide range of approaches have been developed for a range of purposes and regions, classification has become best known and most controversial through its role in guiding stream restoration.

Classification schemes can be seductive, especially for non-geomorphologists, who may not appreciate the complex nature of geomorphological processes and dynamics and who may feel that the channel is completely described and understood once it has been put into a given class. Fitting nature into the classification system may become the objective and important information about the channel may be missed. As the users of a classification system may lack background in fluvial geomorphology, the idealizations of the classification scheme can be more real to them than the evidence presented in the field or the implications that might follow from an understanding of geomorphic processes. In addition, a classification may simply provide a snapshot of channel condition that does not reflect temporal variability, disturbance history or the potential for channel change due to such variability. Perhaps the most important limitation of channel classification is that once put into a class, the channel may be viewed as 'known' and critical thinking abandoned in favour of pre-existing assumptions about this class of channel. The focus of many classifications is on the channel form, not geomorphic processes that control it, largely because the former are easier to quantify.

By focusing on channel form in a reach (at the point in time when the classification is applied), channel classifications cannot capture the dynamic behaviour of river systems or effects of changes upstream and downstream. Notions of 'natural' or reference conditions can be difficult to apply in areas with a long history of human occupation and controlled in some cases by inherited geomorphic features. In small basins, channel form is often linked to nearby geology, land use and basin characteristics, but with increasing drainage area and heterogeneity in the catchment, upstream effects combine with downstream influences on different temporal scales.

In North America, form-based classification is widely used to design channel restoration projects, in lieu of conducting scientifically sound (albeit longer and more costly) geomorphic and ecological studies. The mostly widely used is that of Rosgen (1985, 1994, 1996), a classification scheme based in part on application of hydraulic geometry concepts of Leopold and Maddock (1953) and field experience in the Rocky Mountain region. Rosgen's (1985) scheme recognized 25 distinct stream 'types', based on gradient, sinuosity, width/depth ratio, bed material size and degree of valley confinement, in four major groups along the longitudinal gradient of the river, from steep mountain streams designated 'A', to estuarine channels designated 'D' (Fig. 7.4). Within these four broad groups, specific classes were designated by an alphanumeric code as 'A1', 'A2', 'B1', 'B2', etc. Rosgen modified the channel characteristics attached to each alphanumeric designation several times, recognizing 94 distinct stream classes by 1996 (Rosgen 1996).

Despite strong criticism from the geomorphological community (e.g. Miller and Ritter 1996), the approach has been adopted by various public agencies as offering a standardized approach to prescribe restoration actions (Malakoff 2004; Lave 2008), and it has been institutionalized as required in mitigation projects for wetland impacts in North Carolina (Lave *et al.* 2010). The procedure for applying the system, as indicated by design documents of many such projects, is a determination of the 'proper' stream type for a site based on data collected on-site and expectations of transitions from one type to another. The 'proper' stream type is then constructed with heavy equipment in the expectation that it will be inherently stable. However, to help it remain stable, these projects are typically heavily reinforced on the outside of meander bends with large boulders, rootwads and timbers and the beds with boulder weirs. It is notable that the 'proper' stream type is invariably a single-thread, symmetrically meandering channel, raising interesting questions about unacknowledged cultural preferences for such channel types (Kondolf 2006). This classification system purports to 'predict a river's behaviour from its appearance' (Rosgen 1994, p. 170), although to make valid predictions, the criteria on which the system is based (such as the boundaries separating one 'type' from another) should have geomorphic significance, which has not been demonstrated (Miller and Ritter 1996).

Despite the widespread popularity among practitioners of the Rosgen classification scheme as a basis for restoration design, the actual performance of the projects designed using this approach has been uneven at best, with many failures. For example, a project constructed in 1995 on Deep Run, Maryland, had a stated rationale to reduce bank erosion by tearing out the existing channel and replacing it with a channel whose dimensions would be stable at the site according to the Rosgen classification scheme. The Deep Run project is typical of such projects in its design and fate, but unusual in that it was subject to a thorough post-project appraisal, in which pre-project baseline conditions, as-built project conditions and subsequent performance of the project were documented. Smith and Prestegaard (2005) surveyed pre-project channel geometry and measured flow velocities in the channel and on the floodplain during floods, repeating the surveys and velocity measurements after the project was built. The channel reconstruction project entailed removal of most existing riparian vegetation so that a narrower 'C-4' channel with symmetrical meander bends (based on the Rosgen classification scheme) could be constructed, with channel width and meander amplitude/wavelength based on the designer's estimate of the 'bankfull discharge' (Fig. 7.14). Prior to the project, the channel was highly sinuous, but irregularly so and was flanked by riparian vegetation (Fig. 7.15a). Creating the symmetrical meander bends required that the riparian forest be removed so that the idealized 'C-4' channel could be constructed. After the project was constructed, overbank velocities were higher, presumably owing to the removal of the hydraulically rough riparian vegetation and probably also because the smaller, constructed channel put more water overbank for a given discharge than did the original channel (Smith and Prestegaard 2005). The high overbank velocities allowed Deep Run to cut a new channel through the constructed floodplain, such that within a few years it had largely abandoned the designed channel's bank revetments, outflanking some and eroding others (Figs 7.15b, 7.15c and 7.16).

Figure 7.15 Aerial views of Deep Run, Maryland, prior to and after the 'restoration' project. (a) 1993 pre-project conditions featured an irregularly sinuous channel with riparian forest; (b) same view in 2002, after project construction and failure; (c) same view in 2007 shows continued channel widening. The disturbed area shown in the upper right corner of (a), labelled as 'road construction', is now a completed highway (Maryland Route 100). In addition, the floodplain to the north of Deep Run was by 2007 occupied by a commercial/industrial development. Source: US Geological Survey.

Figure 7.14 Oblique aerial view of the Rosgen-type 'C-4' channel constructed on Deep Run, near Hanover, Maryland, shortly after construction in 1995. Source: Smith and Prestegaard, 2005. Reproduced with permission of Wiley.

The problematic performance of channels such as Deep Run highlights fundamental shortcomings with using a form-based classification system to design channel restoration projects. Although the classification scheme offers an easily applied cookbook approach, it cannot yield the understanding of

Figure 7.16 View downstream on Deep Run in 1999, after project construction, showing root-wad and boulder structures in left bank still intact, but abandoned, as the stream eroded a new channel to the right of the idealized meander bend constructed here. Source: Kondolf, 1999.

fundamental factors leading to channel change (and thus the more sophisticated assessment of opportunities and constraints for restoration for the specific river) that could be gained from geomorphic and ecological studies to understand the current and historical processes in the channel and the catchment. Process-based geomorphic classification systems might perform better, but in the end it is unlikely that any classification system can, by itself, provide an adequate basis for design of restoration projects.

Channel classification: tool or crutch?

Heraclitus's elegant description of the dynamic nature of river channels masks the fact that river channel change may be either perceptible only over centuries or unnervingly rapid, in sudden channel shifts during a single flood. Because the dependent variables of channel geometry (width, depth, slope, velocity and bedform roughness) can adjust in a variety of ways to imposed changes in the independent variables of flow and sediment load, it is impossible to predict with certainty the channel's response to a given perturbation in the system from existing formulae (Maddock 1970; Montgomery and Buffington 1998). Each channel is unique and assessment of its present condition and likely future behaviour require understanding of both its current condition and past behaviour, as documented in historical maps, aerial photographs, surveys and archival sources (Hooke and Kain 1982; Kondolf and Larson 1995; Collins and Montgomery 2001; Collins *et al.* 2002). A historical study should be conducted to determine the stream's characteristic behaviour, especially its response to (and relaxation time from) perturbations such as large floods, changes in sediment supply or engineering works (Chapter 4). Study of fluvial palaeodynamics and temporal trajectories, on scales of multi-centuries to multi-millennia, can also be helpful (Brierley and Fryirs 2005; Buffington and Montgomery 2013).

Systematic description of the existing state of a stream channel can be instructive because the existing channel form integrates the many factors that influence river form and process and a classification may help some managers understand how to describe these existing conditions. However, when using a classification system, it is important to avoid over-emphasizing the categories, lest the user view the stream system as a series of 'snapshots'. Ideally, stream classification should permit the comparison of observations from diverse sites and the application of insights developed in one drainage to another. When used appropriately, channel classification can provide a powerful and flexible tool for fluvial geomorphologists, but the potential to rely on channel classifications as a crutch should warrant substantial care to ensure the judicious use of this tool. It is only when used in this way that it will be useful for restoration and sustainable management of fluvial systems. Certainly for application to river restoration, substantive input from qualified geomorphologists is needed.

Channel classifications are evolving from *a priori* classifications to large geomorphological databases, which can be used flexibly, allowing the end user to build his or her own classification to answer specific questions. With such systems, it is possible to add progressively more data, in addition to historical data and temporal trajectories, to integrate better different time-scales and consider retrospective analysis and to combine corridor and channel information. These databases should integrate GIS software (Gurnell *et al.* 1994) and be used intelligently by well-trained geomorphologists. 'Universal' classifications scheme have only limited usefulness (Kondolf 1995; Heritage *et al.* 1997; Brierley and Fryirs 2000); they cannot be panaceas and cannot replace understanding the channels in question (Montgomery and Buffington 1998; Goodwin 1999). On the other hand, pragmatic applications of channel classification can provide useful tools for understanding smaller geographic units within regional frameworks and broader hierarchical levels (Mosley 1987; Bryce and Clarke 1996; Brierley and Fryirs 2000, 2005; Schmitt *et al.* 2007).

Acknowledgements

This chapter benefited in the first edition from insightful review comments by David Sear and Angela Gurnell. Manuscript preparation was partially supported by the Beatrix Farrand Fund of the Department of Landscape Architecture and Environmental Planning of the University of California, Berkeley.

References

Agence de l'Eau Rhin-Meuse, Laboratoire d'Ecologie de l'Université de Metz, Ecolor, Gereea, Loisirs et d'Étente and INRA Rennes. 1991, *Etude des Végétaux Fixes en Relation avec la Qualité du milieu (Méthodologie MEV)*, Etude Interagences, Hors Série, 94pp.

Alabyan, A.M. and shalov R.S. 1998. Types of river patterns and their natural controls. *Earth Surface Processes and Landforms*. 23: 467–474.

Alber, A. and Piégay, H. 2011. Spatial disaggregation and aggregation procedures for characterizing fluvial features at the network-scale: application to the Rhône basin (France). *Geomorphology* 125: 343–360.

Allan, J.D. and Johnson, L.B. 1997. Catchment-scale analysis of aquatic ecosystems. *Freshwater Biology* 37(1): 107–111.

Amoros, C. and Petts, G.E., eds. 1993. *Hydrosystèmes Fluviaux*. Paris: Masson, 300pp.

Amoros, C., Richardot-Coulet, M. and Pautou, G. 1982. Les 'ensembles fonctionnels': des entités écologiques qui traduisent l'évolution de l'hydrosystème en intégrant la géomorphologie et l'anthropisation (exemple du Haut-Rhône français). *Revue de Géographie de Lyon* 57(1): 49–62.

AQUASCOP. 1997. *Système d'Évaluation de la Qualité du Milieu Physique. 1. Typologie Physique Simplifiée des Cours d'Eau Français. Rapport*, Agences de l'Eau, 55pp.

ASCE (American Society of Civil Engineers, Task Committee on Sediment Transport and Aquatic Habitats). 1992. Sediment and aquatic habitat in river systems. *Journal of Hydraulic Engineering* 118: 669–687.

Astrade, L. and Bravard, J.P. 1999. Energy gradient and geomorphological processes along a river influenced by neotectonics (the Saône river, France). *Geodinamica Acta* 12(1): 1–10.

Beechie, T.J., Liermann, M., Beam,er E.M. and Henderson, R. 2005. A classification of habitat types in a large river and their use by juvenile salmonids. *Transactions of the American Fisheries Society* 134: 717–729.

Beisel, J.-N., Usseglio-Polatera, P., Thomas, S. and Moreteau, J.-C. 1998. Stream community structure in relation to spatial variation: the influence of mesohabitat characteristics. *Hydrobiologia* 389: 73–88.

Belletti, B., Dufour, S. and Piégay, H. 2013. Regional variability of aquatic pattern in braided reaches (example of the French Rhône basin). *Hydrobiologia* 712: 25–41.

Bernot, V. and Creuzé des Châtelliers, D. 1998. *Etude de l'Adaptation de la Méthode de Sectorisation Mise au Point sur le Vidourle à d'Autres Cours d'Eau du Bassin Rhône-Méditerranée-Corse. Rapport Final à l'Agence de l'Eau Rhône-Méditerranée-Corse*, ARALEPBP, 63pp.

Bernot, V., Calland, V., Bravard, J.P. and Foussadier, R. 1996. La sectorisation longitudinale du Vidourle (Gard-Hérault): une méthode appliquée à la gestion environnementale de l'espace fluvial méditerranéen. *Revue de Géographie de Lyon* 4: 323–339.

Bertrand, M., Liébault, F., and Piégay, H. 2013a. Debris-flow susceptibility of upland catchments. *Natural Hazards* 67: 497–511.

Bertrand, M., Piégay, H., Liébault, F., Pont, D., and Sauquet, E. 2013b. Sensitivity analysis of environmental changes associated with riverscape evolutions following sediment reintroduction: application to the Drôme River network, France. *International Journal of River Basin Management*. 11: 19–32.

BES (British Ecological Society). 1990. *Ecological Issues No. 1, River Water Quality*, London:British Ecological Society, Field Studies Council.

Bethemont, J., Andriamahefa, H., Rogers, C. and Wasson, J.G. 1996. Une approche régionale de la typologie morphologique des cours d'eau. Application de la méthode 'morphorégions' au bassin de la Loire et perspectives pour le bassin du Rhône (France). *Revue de Géographie de Lyon*, 71(4): 311–322.

Biggs, B.J.F., Duncan, M.J., Jowett, I.G., Quinn, J.M., Hickey, C.W., Davies-Colley, R.J. and Close, M.E. 1990. Ecological characterisation, classification, and modelling of New Zealand rivers: an introduction and synthesis. *New Zealand Journal of Marine and Freshwater Research* 24: 277–304.

Bostelmann, R., Braukmann, U., Briem, E.U., Drehwald, U.E., Fleischhacker, T.U., Humborg, G.T., Kübler, P.G., Nadolny, I. and Scheurlen, K. 1998a. Regionale Bachtypen in Baden-Würtemberg. Arbeitsweisen und exemplarische Ergebnisse an Keuper- und Gneisbächen. In: *Handbuch Wasser 2*, Karlsruhe: LFU, Ministerium für Umwelt undVerkehr Baden-Würtemberg, 268pp.

Bostelmann, R., Brauckmann, U., Briem, E., Fleischhacker, T., Humborg, G., Nadolny, I., Scheurlen, K. and Weibel, U. 1998b. An approach to classification of natural streams and floodplains in south-west Germany. In: Waal, L.C., Large, A.R.G., and Wade, P.M., eds., *Rehabilitation of Rivers: Principles and Implementation*, Chichester: John Wiley & Sons, pp. 31–55.

Bravard, J.P. and Gilvear, D.J. 1993. Structure hydro-géomorphologique des hydrosystèmes. In: Amoros, C. and Petts, G.E., eds., *Hydrosystèmes Fluviaux. Collection d'Écologie*, vol. 24, Paris: Masson, pp. 83–103.

Bravard, J.P. and Peiry, J.L. 1999. The *CM* pattern as a tool for the classification of alluvial suites and floodplains along the river continuum. In: Marriott, S.B. and Alexander, J., eds., *Floodplains: Interdisciplinary Approaches, Special Publication 163*, London: Geological Society, pp. 259–268.

Bravard, J.P., Amoros, C. and Pautou, G. 1986. Impact of civil engineering works on the successions of communities in a fluvial system. A methodological and predictive approach applied to a section of the Upper Rhône River, France. *OIKOS* 47: 92–111.

Brice, J.C. 1982. *Stream Channel Stability Assessment*. Technical Report No. FHWA/RD-82/021, Washington, DC: Federal Highways Administration, US Department of Transportation.

Brierley, G.J. and Fryirs, K. 2000. River styles, a geomorphic approach to catchment characterization: implications for river rehabilitation in Bega catchment, New South Wales, Australia. *Environmental Management* 25(6): 661–679.

Brierley, G.J., and Fryirs, K.A. 2005. *Geomorphology and River Management: Applications of the River Styles Framework*, Oxford: Blackwell.

Brookes, A. 1987. The distribution and management of channelized streams in Denmark. *Regulated Rivers – Research and Management* 1: 3–16.

Brookes, A. 1990. Restoration and enhancement of engineered river channels: some European experiences. *Regulated Rivers – Research and Management* 5: 45–56.

Brown A.G. 2002. Learning from the past: palaeohydrology and palaeoecology. *Freshwater Biology* 47: 817–829.

Brussock, P.P., Brown, A.V. and Dixon, J.C. 1985. Channel form and stream ecosystem models. *Water Resources Bulletin* 21: 859–866.

Bryce, S.A. and Clarke, S.E. 1996. Landscape-level ecological regions: linking state-level ecoregion frameworks with stream habitat classifications. *Environmental Management* 20(3): 297–311.

Buffington, J.M. and Montgomery, D.R. 2013. Geomorphic classification of rivers. In: Shroder, J. and Wohl, E., eds., *Treatise onGeomorphology, Vol. 9, Fluvial Geomorphology*, San Diego, CA: Academic Press, pp. 730–767.

Carbiener, R. 1983. Le grand Ried Central d'Alsace: écologie et évolution d'une zone humide d'origine fluviale rhénane. *Bulletin d'Ecolologie* 14: 249–277.

Carpenter, K.E. 1928. *Life in Inland Waters with Especial Reference to Animals*, London: Sidgwick and Jackson (as cited by Hawkes 1975).

Castella, E., Béguin, O., Besacier-Monbertrand, A.-L., Hug Peter, D., Lamouroux, N., Mayor Siméant, H., McCrae, D., Olivier, J.-M. and Paillex, A. 2015. Realised and predicted changes in the invertebrate benthos after restoration of connectivity to the floodplain of a large river. *Freshwater Biology* 60: 1131–1146.

Chandesris, A., Mengin, N., Malavoi, J.R., Souchon, Y. and Wasson, J.G. 2009. *Système Relationnel d'Audit de l'Hydromorphologie des Cours d'Eau (SYRAH_CE), Atlas à Large Échelle*, vol. 2, Lyon, online report, 58pp.

Chessman, B.C., Fryirs, K.A. and Brierley, G.J. 2006. Linking geomorphic character, behaviour and condition to fluvial biodiversity: implications for river management. *Aquatic Conservation: Marine and Freshwater Ecosystems* 16: 267–288.

Church, M. 1992. Channel morphology and typology. In: Calow, P. and Petts, G.G., eds., *The Rivers Handbook: Hydrological and Ecological Principles*, Oxford: Blackwell, pp. 126–143.

Church, M. 2006. Bed material transport and the morphology of alluvial rivers. *Annual Review of Earth and Planetary Sciences* 34: 325–354.

Collins, B.D. and Montgomery, D.R. 2001. Importance of archival and process studies to characterizing pre-settlement riverine geomorphic processes and habitat in the Puget Lowland. In: Dorava, J.B., Montgomery, D.R., Palcsak, B. and Fitzpatrick, F., eds., *Geomorphic Processes and Riverine Habitat*, Washington, DC: American Geophysical Union, pp. 227–243.

Collins, B.D., Montgomery, D.R., and Haas, A. 2002. Historic changes in the distribution and functions of large woody debris in Puget Lowland rivers. *Canadian Journal of Fisheries and Aquatic Sciences* 59: 66–76.

Corbonnois, J. and Zumstein, J.F. 1994. Proposition de typologie des cours d'eau. Application au réseau hydrographique du Nord-Est de la France (bassin de la Moselle). *Revue de Géographie Alpine* 2: 15–24.

Cupp, C.E. 1989a. *Identifying Spatial Variability of Stream Characteristics Through Classification*, MS Thesis, University of Washington, Seattle, WA, 92 pp.

Cupp, C.E. 1989b. *Stream Corridor Classification for Forested Lands of Washington*, Olympia, WA: Washington Forest Protection Association, 24pp.

Curran, J.C. and Wilcock, P.R. 2005. Characteristic dimensions of the step-pool bed configuration: an experimental study. *Water Resources Research* 41: W02030.

Cushing, C.E., McIntire, C.D., Cummins, K.W., Minshall, G.W., Petersen, R.C., Sedell, J.R. and Vannote, R.L. 1983. Relationships among chemical, physical, and biological indices along river continua based on multivariate analyses. *Archiv für Hydrobiologie* 98: 317–326.

Dana, J. 1850. On denudation in the Pacific. *American Journal of Science, Ser*. 2 9: 48–62.

Davis, W.M. 1899. The geographical cycle. *Geographical Journal* 14: 481–504.

Downs, P.W. 1994. Characterization of river channel adjustements in the Thames basin, South-East England. *Regulated Rivers – Research and Managment* 9: 151–175.

Downs, P.W. 1995. Estimating the probability of river channel adjustment. *Earth Surface Processes and Landforms* 20: 687–705.

Downs, P.W. and Brookes, A. 1994. Developing a standard geomorphological approach for the appraisal of river projects. In: Kirkby, C. and White, W.R., eds., *Integrated River Basin Development*, Chichester: John Wiley & Sons, pp. 299–310.

Doyle, M.W., Miller, D.E. and Harbor, J.M. 2000. Should river restoration be based on classification schemes or process models? Insights from the history of geomorphology. In: Walton, R. and Nece, R.E., eds., *Proceedings of the ASCE International Conference on Water Resources Engineering, Seattle, 1999*, Reston, VA: American Society of Civil Engineers.

Dufour, S. and Piégay, H. 2009. From the myth of a lost paradise to targeted river restoration: forget natural references and focus on human benefits. *River Research and Applications* 25: 568–581.

Emery, J.C., Gurnell, A.M., Clifford, N.J., Petts, G.E., Morrissey, I.P. and Soar, P.J. 2003. Classifying the hydraulic performance of riffle-pool beforms for habitat assessment and river rehabilitation design. *River Research and Applications* 19: 533–549.

Environment Agency. 2002. *RHS Applications Manual Version 1.1*, Bristol: Environment Agency, 71pp.

Erba, S., Buffagni, A., Holmes, N., O'Hare, M., Scarlett, P. and Stenico, A. 2006. Preliminary testing of River Habitat Survey features for the aims of the WFD hydro-morphological assessment: an overview from the STAR Project. *Hydrobiologia* 566: 281–296.

European Parliament and Council of Europe. 2000. Directive 2000/60/EC of the European Parliament and of the Council of

23 October 2000 establishing a framework for Community action in the field of water policy. *Official Journal* L 327(22/12/2000): 1–73.

Ferguson, R. 1981. Channel form and channel changes. In: Lewin, J., ed., *British Rivers*, London: Allen & Unwin, pp. 90–125.

Ferguson, R. 1987. Hydraulic and sedimentary controls of channel pattern. In: Richards, K., ed., *River Channels. Environment and Process*, Oxford: Blackwell, pp. 129–158.

Ferguson, R.J. and Brierley, G.J. 1999. Downstream changes in valley confinement as a control on floodplain morphology, Lower Tuross River, New South Wales, Australia: a constructivist approach to floodplain analysis. In: Miller, A.J. and Gupta, A., eds., *Varieties of Fluvial Form*, Chichester: John Wiley & Sons, pp. 377–407.

Ferréol, M., Dohet, A., Cauchie, H.-M. and Hoffmann, L. 1998. An environmental typology of freshwater sites in Luxembourg as a tool for predicting macroinvertebrate fauna under non-polluted conditions. *Ecological Modelling* 212: 99–108.

Frissell, C.A., Liss, W.J., Warren, C.E. and Hurley, M.D. 1986. A hierarchical framework for stream habitat classification: viewing streams in a watershed context. *Environmental Management* 10: 199–214.

Furse, M.T., Moss, D., Wright, J.F. and Armitage, P.D. 1984. The influence of seasonal and taxonomic factors on the ordination and classification of running-water sites in Great Britain and on the prediction of their macro-invertebrate communities. *Freshwater Biology* 14: 257–280.

Galay, V.J., Kellerhals, R. and Bray, D.I. 1973. Diversity of river types in Canada. In: *Fluvial Processes and Sedimentation, Proceedings of Hydrology Symposium No. 9*, Edmonton, Alberta: National Research Council of Canada, pp. 216–250.

Gilbert, G.K. 1877. *Report on the Geology of the Henry Mountains, U.S. Geographical and Geological Survey of the Rocky Mountain Region*, Washington, DC: US Government Printing Office.

Goodwin, C.N. 1999. Fluvial classification: Neanderthal necessity or needless normalcy? In: Olson, D.S., and Potyondy, J.P., eds., *Wildland Hydrology*, Herndon, VA: American Water Resources Association, pp. 229–236.

Grant, G.E., Swanson, F.J. and Wolman, M.G. 1990. Pattern and origin of stepped-bed morphology in high-gradient streams, Western Cascades, Oregon. *Geological Society of America Bulletin* 102: 340–352.

Gregory, S. and Ashkenas, L. 1990. *Riparian Management Guide: Willamette National Forest*, Eugene, OR: US Department of Agriculture, Forest Service, Pacific Northwest Region.

Grosprêtre, L. 2011. *Etude et Gestion des Impacts Hydrogéomorphologiques de la Périurbanisation. L'exemple du Bassin de l'Yzeron dans l'Ouest Lyonnais*, PhD Thesis, University of Lyon 2, 280pp.

Gurnell, A.M., Angold, P. and Gregory, K.J. 1994. Classification of river corridors: issues to be adressed in developing an operational methodology. *Aquatic Conservation: Marine and Freshwater Ecosystems* 4: 219–231.

Gustard, A. 1992. Analysis of river regimes. In: Calow, P. and Petts, G.E., eds., *The Rivers Handbook. Hydrological and Ecological Principles*, vol. 1, Oxford: Blackwell, pp. 29–47.

Hallot, E. 2010. *Typologie Hydro-géomorphologique des Cours d'Eau dans l'Euroregio Meuse-Rhin*, PhD Thesis, Académie Universitaire Wallonie-Europe, Liège University, 273pp.

Harper, D. and Everard, M. 1998. Why should the habitat level approach underpin holistic river survey and management? *Aquatic Conservation: Marine and Freshwater Ecosystems* 8: 395–413.

Harvey, G.L., Clifford, N.J. and Gurnell, A.M.. 2008. Towards an ecologically meaningful classification of the flow biotope for river inventory, rehabilitation, design and appraisal purposes. *Journal of Environmental Management* 88: 638–650.

Hawkes, H.A. 1975. River zonation and classification. In: Whitton, B., ed., *River Ecology*,. Berkeley, CA: University of California Press, pp. 312–374.

Hawkins, C.P., Kershner, J.L., Bisson, P.A., Bryant, M.D., Decker, L.M., Gregory, S.V., et al. 1993. A hierarchical approach to classifying stream habitat features. *Fisheries* 18: 3–12.

Henry, C.P. and Amoros, C. 1995. Restoration ecology of riverine wetlands: I. A scientific base. *Environmental Management* 19: 891–902.

Heritage, G.L., Van Niekerk, A.W. and Moon, B.P. 1997. A comprehensive hierarchical river classification system. *Geoökoplus* 4: 75–84.

Hicks, D.M. and Mason, P.D. 1991. *Roughness Characteristics of New Zealand Rivers*. Wellington: New Zealand Department of Scientific and Industrial Research, Marine and Freshwater, Water Resources Survey.

Holmes, N. 1989. British rivers: a working classification. *British Wildlife* 1: 20–36.

Hooke, J.M. and Kain, R.J.P. 1982. *Historical Change in the Physical Environment: a Guide to Sources and Techniques*, London: Butterworth.

Horton, R.E. 1945. Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology. *Bulletin of the Geological Society of America* 56: 275–370.

Huet, M. 1949. Aperçu des relations entre la pente et les populations piscicoles des eaux courantes. *Schweizerische Zeitschrift für Hydrologie* 11(3–4): 332–351.

Hugues, R.M., Heiskary, S.A., Matthews, W.J. and Yoder, C.O. 1993. Use of ecoregions in biological monitoring. In: Loeb, S.L. and Spacie, A., eds., *Biological Monitoring of Aquatic Systems*, Boca Raton, FL: Lewis Publishers, pp. 125–151.

Hynes, H.B.N. 1975. Edgardo Baldi Memorial Lecture. *The stream and its valley. Verhandlungen der Internationalen Vereinigung für Theoretische und Angewandte Limnologie* 19: 1–15.

Imhof, J.G., Fitzgibbon, J. and Annable, W.K. 1996. A hierarchical evaluation system for characterizing watershed ecosystems for fish habitat. *Canadian Journal of Fisheries and Aquatic Sciences* 53(Suppl. 1): 312–326.

Jeffers, J.N.R. 1998. Characterization of river habitats and prediction of habitat features using ordination techniques. *Aquatic Conservation: Marine and Freshwater Ecosystems* 8: 529–540.

Johnson, D.W. 1944. Problems of terrace correlation. *Bulletin of the Geological Society of America* 55: 793–818.

Jowett, I.G. and Duncan, M.J. 1990. Flow variability in New Zealand rivers and its relationship to in-stream habitat and biota. *New Zealand Journal of Marine and Freshwater Research* 24: 305–317.

Kellerhals, R., Church, M. and Bray, D.I. 1976. Classification and analysis of river processes. *Journal of the Hydraulics Division, American Society of Civil Engineers* 102: 813–829.

Kern, K. 1992. Restoration of lowland rivers: the German experience. In: Carling, P.A. and Petts, G.E., eds., *Lowland Floodplain Rivers: Geomorphological Perspectives*, Chichester: John Wiley & Sons, pp. 279–297.

Kern, K. 1994. *Grundlagen Naturnaher Gewassergestaltung-Geomorphologische Entwicklung von Fliessgewassern*, Berlin: Springer, 256pp.

King, J.M., DeMoor, F.C. and Chutter, F.M. 1992. Alternative ways of classifying rivers in southern Africa. In: Boon, P.J., Calow, P. and Petts, G.E., eds., *River Conservation and Management*, Chichester: John Wiley & Sons, pp. 213–230.

Knighton, A.D. 1999. Downstream variation in stream power. *Geomorphology* 29(3–4): 293–306.

Knighton, A.D. and Nanson, G.C. 1993. Anastomosis and the continuum of channel pattern. *Earth Surface Processes and Landforms* 18: 613–625.

Kondolf, G.M. 1995. Geomorphological stream channel classification in aquatic habitat restoration: uses and limitations. *Aquatic Conservation* 5: 127–141.

Kondolf, G.M. 2006. River restoration and meanders. *Ecology and Society* 11(2): 42 [online]: http://www.ecologyandsociety.org/vol11/iss2/art42/.

Kondolf, G.M. and Larson, M. 1995. Historical channel analysis and its application to riparian and aquatic habitat restoration. *Aquatic Conservation* 5: 109–126.

Lave, R. 2008. *The Rosgen Wars and the Shifting Political Economy of Expertise*, PhD Thesis, University of California, Berkeley, 251pp.

Lave, R., Doyle, M. and Robertson, M. 2010. Privatizing stream restoration in the US. *Social Studies of Science* 40: 677–703.

Leopold, L.B. and Maddock, T. 1953. *The Hydraulic Geometry of Stream Channels and Some Physiographic Implications, US Geological Survey Professional Paper 252*, Washington, DC: US Government Printing Office, 57pp.

Leopold, L.B. and Wolman, M.G. 1957. *River Channel Patterns – Braided, Meandering, and Straight, US Geological Survey Professional Paper 282(B)*, Washington, DC: US Government Printing Office, pp. 39–85.

Leopold, L.B., Wolman, M.G. and Miller, J.P. 1964. *Fluvial Processes in Geomorphology*, San Francisco: W.H. Freeman.

Leviandier, T., Alber, A., Le Ber, F. and Piégay, H. 2012. Comparison of statistical algorithms for detecting homogeneous river reaches along a longitudinal continuum. *Geomorphology* 138, 130–144.

Lotspeich, F.B. 1980. Watersheds as the basic ecosystem: this conceptual framework provides a basis for a natural classification system. *Water Ressources Bulletin* 16(4): 581–586.

Maddock, T. Jr., 1970. Indeterminate hydraulics of alluvial channels. *Journal of the Hydraulics Division, American Society of Civil Engineers* 96: 2309–2323.

Maire, G. and Wilms, P. 1984. *Etude Hydro-gémorphologique du Giessen. Détermination de Secteurs Homogènes*. Strasbourg: Région Alsace, Université Louis Pasteur, Centre de Géographie Appliquée, UA 95 du CNRS, 78pp.

Makaske, B. 2001. Anastomosing rivers: a review of their classification, origin and sedimentary products. *Earth Sciences Reviews* 53: 149–196.

Malakoff, D. 2004. The river doctor. *Science* 305: 937–939.

Malavoi, J.R. 1989. Typologie des faciès d'écoulement ou unités morphodynamiques des cours d'eau à haute énergie. *Bulletin Français de Pêche et de Pisciculture* 315: 189–210.

Malavoi J.R. 2000. *Typologie et Sectorisation des Cours d'Eau du Bassin Loire-Bretagne*, Orléans: Agence de l'Eau Loire-Bretagne, 79pp.

Miller, J.R. and Ritter, J.B. 1996. Discussion. An examination of the Rosgen classification of natural rivers. *Catena* 27: 295–299.

Mollard, J.D. 1973. Air photo interpretation of fluvial features. In: *Fluvial Processes and Sedimentation. Proceedings of Canadian Hydrology Symposium, Edmonton, Alberta, 1973*, Ottawa: National Research Council of Canada, Subcommittee on Hydrology, pp. 341–380.

Montgomery, D.R. 1999. Process domains and the river continuum. *Journal of the American Water Resources Association* 35: 397–410.

Montgomery, D.R. and Buffington, J.M. 1997. Channel reach morphology in mountain drainage basins. *Geological Society of America Bulletin* 109: 596–611.

Montgomery, D.R. and Buffington, J.M. 1998. Channel processes, classification, and response potential. In: Naiman, R.J., and Bilby, R.E., eds., *River Ecology and Management*, New York: Springer, pp. 13–42.

Montgomery, D.R. and Gran, K.B. 2001. Downstream hydraulic geometry of bedrock channels. *Water Resources Research* 37: 1841–1846.

Montgomery, D.R., Dietrich, W.E. and Sullivan, K. 1998. The role of GIS in watershed analysis. In: Lane, S.N., Richards, K.S. and Chandler, J.H., eds., *Landform Monitoring, Modelling and Analysis*, Chichester: John Wiley and Sons, pp. 241–261.

Morandi, B., Piégay, H., Lamouroux, N. and Vaudor, L. 2014. How is success or failure in river restoration projects evaluated? Feedback from French restoration projects. *Journal of Environmental Management* 137: 178–188.

Mosley, M.P. 1982. *A Procedure for Characterising River Channels, Water and Soil Miscelleneous Publication No. 32*, Christchurch: New Zealand Ministry of Works and Development.

Mosley, M.P. 1987. The classification and characterization of rivers. In: Richards, K., ed., *River Channels, Environment and Process*, Oxford: Blackwell, pp. 295–320.

Müller, A., Glacer, D., Sommerhäuser, M. and Timm, T. 1996. Leitbilder für die Gewässerstrukturgütekartierung in Nordrhein-Westfalen. *Kasseler Wasserbau-Mitteilungen* 6: 95–105.

Naiman, R.J., Lonzarich, D.G., Beechie, T.J. and Ralph, S.C. 1992. General principles of classification and the assessment of conservation potential in rivers. In: Boon, P.J, Calow, P. and Petts, G.E., eds., *River Conservation and Management*, Chichester: John Wiley & Sons, pp. 93–124.

Nanson, G.C. and Croke, J.C. 1992. A genetic classification of floodplains. *Geomorphology* 4: 459–486.

Nanson, G.C. and Knighton, A.D. 1996. Anabranching rivers: their causes, character and classification. *Earth Surface Processes and Landforms* 21(3): 217–239.

Nevins, T.H.F. 1965. River classification with particular respect to New Zealand. In: *Proceedings of 4th New Zealand Geography Conference, Dunedin. New Zealand, Geographical Society Conference Series No. 4*, Palmerston North: New Zealand Geographical Society, pp. 83–90.

Newson, M.D. and Newson, C.L. 2000. Geomorphology, ecology and river channel habitat: mesoscale approaches to basin-scale challenges. *Progress in Physical Geography* 24(2): 195–217.

Newson, M.D., Clark, M.J., Sear, D.A. and Brookes, A.. 1998. The geomorphological basis for classifying rivers. *Aquatic Conservation* 8: 415–430.

Nicolas, Y. and Pont, D. 1997. Hydrosedimentary classification of natural and engineered backwaters of a large river, the lower Rhône: possible applications for the maintenance of high fish biodiversity. *Regulated Rivers – Research and Management* 13: 417–431.

Notebaert, B.and Piégay, H. 2013. Multi-scale factors controlling the pattern of floodplain width at a network scale: the case of the Rhône basin, France. *Geomorphology* 200: 155–171.

NRA (National Rivers Authority) 1992. *River Channel Typology for River Planning and Management, Internal Report to NRA Thames Region, prepared by Geodata Unit*, Southampton University.

Ollero, O.A., Ballarin, F.D., Diaz, B.E., Mora, M.D., Sanchez, F.M., Acin, N.V., et al. 2007. Un indice hidrogeomorfologico (IHG) para la evaluacion del estado ecologico de sistemas fluviales. *Geographicalia* 52: 113–141.

Ollero, A., Ibisate, A., Gonzalo, L.E., Acín, V., Ballarín, D., Díaz, E., et al. 2011. The IHG index for hydromorphological quality assessment of rivers and streams: updated version. *Limnetica* 30(2): 255–262.

Omernik, J.M. 1987. Ecoregions of the coterminous United States. *Annals of the Association of American Geographers* 77: 118–125.

Orlowski, L.A., Schumm, S.A. and Mielke, P.W. Jr., 1995. Reach classifications of the lower Mississippi River. *Geomorphology* 14: 221–234.

Orr, H.G., Large, A.R.G., Newson, M.D. and Walsh, C.L.. 2008. A predictive typology for characterising hydromorphology. *Geomorphology* 100: 32–40.

Otto, A. 1991. Grundlagen einer morphologischen Typologie der Bache. In: Larsen, P., ed., *Mitteilungen 180*, Karlsuhe: University of Karlsruhe, Institute for Hydraulic Structures and Agricultural Engineering, pp. 1–94.

Otto, A. and Braukmann, U. 1983. Gewässertypologie im ländlichen Raum. In: *Schriftenreihe des Bundesministers für Ernährung, Landwirtschaft und Forsten. Reihe A: Angewandte Wissenschaft, Heft 288*, Münster: Landwirtschaftsverlag GmbH, 61pp.

Palmer, L. 1976. River management criteria for Oregon and Washington. In: Coates, D.R., ed., *Geomorphology and Engineering*, Stroudsburg, PA: Dowden, Hutchinson, and Ross, pp. 329–346.

Paustian, S.J., Marion, D.A. and Kelliher, D.F. 1984. Stream channel classification using large scale aerial photography for Southeast Alaska watershed management. In: Murtha, P.A. and Harding, R.A., eds., *Renewable Resources Management: Application of Remote Sensing. Proceedings of the RNRF Symposium on the Application of Remote Sensing to Resource Management, Seattle, WA, May 22–27, 1983*, Falls Church, VA: American Society of Photogrammetry, pp. 670–677.

Paustian, S.J., et al. 1992. *A Channel Type Users Guide for the Tsongas National Forest, Southeast Alaska, Technical Paper 26*, Juneau, AL: USDA Forest Service, Alaska Region.

Pennak, R.W. 1971. Towards a classification of lotic habitats. *Hydrobiologia* 38: 321–334.

Petit, F. 1995. Régime hydrologique et dynamique fluviale des rivières ardennaises. In: Demoulin, A., ed., *L'Ardenne, Essai de Géographie Physique, Livre en Hommage au Professeur A. Pissart*, Liège: Laboratoire d'Hydrographie et de Géomorphologie Fluviatile, Université de Liège, pp. 194–223.

Piégay, H., Barge, O. and Landon, N. 1996. Streamway concept applied to river mobility/human use conflict management. In: Maxwell, W.H.C., Preul, H.C. and Stout, G.E., eds., *Rivertech '96: 1st International Conference on New/Emerging Concept for Rivers, Chicago, USA, September 22–26, 1996*, Montpellier: International Water Resources Association, pp. 681–688.

Piégay, H., Cuaz, M., Javelle, E. and Mandier, P. 1997. A new approach to bank erosion management: the case of the Galaure River, France. *Regulated Rivers – Research and Management* 13: 433–448.

Piégay, H., Salvador, P.G. and Astrade, L.. 2000. Réflexions relatives à la variabilité spatiale de la mosaïque fluviale à l'échelle d'un tronçon. *Zeitschrift für Geomorphologie* 44(3): 317–342.

Piégay, H., Darby, S.E., Mosselman, E. and Surian, N. 2005. The erodible corridor concept: applicability and limitations for river management. *River Research and Applications* 21: 773–789.

Piégay, H., Naylor, L.A., Haidvogl, G., Kail, J., Schmitt, L. and Bourdin, L. 2008. Integrative river science and rehabilitation: European experiences. In: Brierley, G. and Fryirs, K., eds., *River Futures. An Integrative Scientific Approach to River Repair*, Washington, DC: Island Press, pp. 201–219.

Piégay, H., Alber, A., Slater, L. and Bourdin, L. 2009. Census and typology of braided rivers in the French Alps. *Aquatic Sciences* 71: 371–388.

Piégay, H., Alber, A., Lauer, J., Rollet, A.-J. and Wiederkehr, E. 2012. Biophysical characterisation of fluvial corridors at reach to network scales. In: Carbonneau, P. and Piégay, H., eds., *Fluvial Remote Sensing for Science and Management*, Chichester: John Wiley & Sons, pp. 241–269.

Platts, W.S. 1980. A plea for fishery habitat classification. *Fisheries* 5: 2–6.

Platts, W.S., Megahan, W.F. and Minshall, G.W. 1983. *Methods for Evaluating Stream, Riparian, and Biotic Conditions, USDA Forest Service General Technical Report INT-138*, Ogden, UT: USDA Forest Service Intermountain Research Station.

Platts, W.S., Armour, C., Booth, G.D.C., Bryant, M., Cuplin, P., Jensen, S., et al. 1987. *Methods for Evaluating Riparian Habitats with Applications to Management, USDA Forest Service General Technical Report INT-221*, Ogden, UT: USDA Forest Service Intermountain Research Station.

Pont, D., Piégay, H., Farinetti, A., Allain, S., Landon, N., Liébault, F., Dumont, B. and Richard-Mazet A. 2009. Conceptual framework and interdisciplinary approach for the sustainable management of gravel-bed rivers: the case of the Drôme River basin (SE France). *Aquatic Sciencess* 71(3): 356–370.

Powell, J.W. 1875. *Exploration of the Colorado River of the West and Its Tributaries*, Washington, DC, US Government Printing Office, 291pp.

Raven, P.J., Fox, P., Everard, M.P., Holmes, N.T.H. and Dawson, F.H. 1997. River Habitat Survey: a new system for classifying rivers according to their habitat quality. In: Boon, P.J. and Howell, D.L., eds., *Freshwater Quality: Defining the Indefinable?, Scottish Natural Heritage*, Edinburgh: The Stationery Office, pp. 215–234.

Raven, P.J., Boon, P.J., Dawson, F.H. and Ferguson, A.J.D. 1998. Towards an integrated approach to classifying and evaluating rivers in the UK. *Aquatic Conservation: Marine and Freshwater Ecosystems* 8: 383–393.

Raven, P.J., Holmes, N.T.H., Charrier, P., Dawson F.H., Naura, M. and Boon, P.J. 2002. Towards a harmonized approach for hydromorphological assessment of rivers in Europe: a qualitative comparison of three survey methods. *Aquatic Conservation: Marine and Freshwater Ecosystems* 12: 405–424.

Rinaldi, M., Surian, N., Comiti, F. and Bussettini, M. 2013. A method for the assessment and analysis of the hydromorphological condition of Italian streams: the Morphological Quality Index (MQI). *Geomorphology* 180–181: 96–108.

Riquier, J., Piégay, H. and Šulc Michalková, M. 2015. Hydromorphological conditions in eighteen restored floodplain channels of a large river: linking patterns to processes. *Freshwater Biology* 60: 1085–1103.

Robach, F., Thiébaut, G., Trémolières, M. and Muller, S. 1996. A reference system for continental running waters: plant communities as bioindicators of increasing eutrophication in alkaline and acidic waters in north-east France. *Hydrobiologia* 340: 67–76.

Rohm, C.M., Giese, J.W. and Bennett, C.C. 1987. Evaluation of an aquatic ecoregion classification of streams in Arkansas. *Journal of Freschwater Ecology* 4(1): 127–140.

Rosgen, D.L. 1985. A stream classification system. In: *Riparian Ecosystems and Their Management, Proceedings of the First North American*

Riparian Conference, April 16–18, Tucson, Arizona, General Technical Report No. RM-120, Fort Collins, CO: USDA Forest Service pp. 91–95.

Rosgen, D.L. 1994. A classification of natural rivers. *Catena* 22(3): 169–199.

Rosgen, D.L. 1996. *Applied River Morphology*, Pagosa Springs, CO: Wildland Hydrology, 390pp.

Roux, C., Alber, A., Bertrand, M., Vaudor, L. and Piégay, H. 2015. 'FluvialCorridor': a new ArcGIS toolbox package for multiscale riverscape exploration. *Geomorphology* 242: 29–37.

Rust, B.R. 1978. A classification of alluvial channel systems. In: Miall, A.D., ed., *Fluvial Sedimentology, Canadian Society of Petroleum Geologists Memoir 5*, Calgary: Canadian Society of Petroleum Geologists, pp. 187–198.

Schmitt, L. 2001. *Typologie Hydro-géomorphologique Fonctionnelle de Cours d'Eau. Recherche Méthodologique Appliquée aux Systèmes Fluviaux d'Alsace*, PhD Thesis, University of Strasbourg, 217pp.

Schmitt, L., Maire, G. and Humbert, J. 2000. Typologie hydro-géomorphologique des cours d'eau : vers un modèle adapté à la gestion du milieu physique des rivières du versant sud-occidental du fossé rhénan. *GéoCarrefour* 75: 347–363.

Schmitt, L., Maire, G. and Humbert, J. 2001. La puissance fluviale: définition, intérêt et limites pour une typologie hydro-géomorphologique de rivières. *Zeitschrift für Geomorphologie* 45: 201–224.

Schmitt, L., Valette, L., Valin, K., Piégay, H. and Hallot, E. 2004. Proposition d'une méthode de typologie hydro-géomorphologique des cours d'eau et test sur un sous-bassin du Rhône (bassin de l'Yzeron). *Mosella* 29(3–4): 323–340.

Schmitt, L., Maire, G., Nobelis, P. and Humbert, J. 2007. Quantitative morphodynamic typology of rivers. A methodological study based on the French Upper Rhine basin. *Earth Surface Processes and Landforms* 32(11): 1726–1746.

Schmitt, L., Lafont, M., Trèmolières, M., Jezequel, C., Vivier, A., Breil, P., Namour, P., Valin, K. and Valette, L. 2011. Using hydro-geomorphological typologies in functional ecology: preliminary results in contrasted hydrosystems. *Physics and Chemistry of the Earth* 36: 539–548.

Schumm, S.A. 1963. *A Tentative Classification of Alluvial River Channels. US Geological Survey Circular 477*, Washington, DC: US Geological Survey.

Schumm, S.A. 1977. *The Fluvial System*, New York: John Wiley & Sons.

Schumm, S.A. 1985. Patterns of alluvial rivers. *Annual Review of Earth and Planetary Sciences* 13: 5–27.

Simons, P.K. 1978. The river scene: comments on North Island rivers. In: *Proceedings of Erosion Assessment and Control Conference, Christchurch, New Zealand*, Christchurch: New Zealand Association of Soil Conservators, pp. 156–158.

Smith, D.G. and Smith, N.D. 1980. Sedimentation in anastomosed river systems: examples from alluvial valleys near Banff, Alberta. *Journal of Sedimentary Petrology* 50(1): 157–164.

Smith, S.M., and Prestegaard, K.L.. 2005. Hydraulic performance of a morphology-based stream channel design. *Water Resources Research* 41: W11413.

Sneath, P.H.A. and Snokal, R.R. 1973. *Numerical Taxonomy: the Principles and Practice of Numerical Classification*, San Francisco: W.H. Freeman,.

Souchon, Y., Andriamahefa, H., Cohen, P., Breil, P., Pella, H., Lamouroux, N., Malavoi, J.R. and Wasson, J.B. 2000. *Régionalisation de l'Habitat dans le Bassin de la Loire, unpublished report*, Cemagref, Agence de l'Eau Loire Bretagne, p. 291.

Strahler, A.N. 1957. Quantitative analysis of watershed geomorphology. *Transactions, American Geophysical Union* 38(6): 913–920.

Terrell, T.T. and McConnell, W.J. 1978. In: *Stream Classification 1977: Proceedings of a Workshop held at Pingree Park, Colorado, October 1977, US Fish and Wildlife Service Publication No. FWS/OBS-78/23*, Fort Collins, CO: USFWS Instream Flow Group, pp. 1–38.

Thorne, C.R. 1997. Channel types and morphological classification. In: Thorne, C.R., Hey, R.D. and Newson, M.D., eds., *Applied Fluvial Geomorphology for River Engineering and Management*, Chichester: John Wiley & Sons, pp. 175–222.

Thorne, C.R. 1998. *Stream Reconnaissance Handbook*, Chichester: John Wiley & Sons, 133pp.

Tölk, J. 1998. *Leitbilderstellung für fliessgewässer im Kraichgau*, Diplomarbeit, Institut für Geographie und Geoökologie der Universität Karlsruhe, Gewässerdirektion Nördlicher Oberrhein, 138pp..

Tricart, J. 1977. *Précis de Géomorphologie. Tome 2, Géomorphologie Dynamique Générale*. Paris: SEDES, 345pp.

Van den Berg, J.H. 1995. Prediction of alluvial channel pattern of perennial rivers. *Geomorphology* 12: 259–279.

Vannote, R.L., Minshall, G.W., Cummins, K.W., Sedell, J.R. and Cushing, C.E.C. 1980. The river continuum concept. *Canadian Journal of Fisheries and Aquatic Sciences* 37: 130–137.

Vocal Ferencevic, M. and Ashmore, P. 2012. Creating and evaluating digital elevation model-based stream-power map as stream assessment tool. *River Research and Applications* 28: 1394–1416.

Wasson, J.G., Bethemont, J., Degorce, J.N., Dupuis, B. and Joliveau, T. 1993. *Approcheé Cosystèmique du Bassin de la Loire: Eléments pour l'Élaboration des Orientations Fondamentales de Gestion. Phase 1: Etat Initial – Problématique*, Rapport d'Étape, CEMAGREF, Groupement de Lyon et Université Jean Monnet de St. Etienne, CRENAM, CNRS URA 260, 102pp.

Wawrzyniak, V., Piégay, H., Allemand, P., Vaudor, L. and Grandjean, P. 2013. Prediction of water temperature heterogeneity of braided rivers using very high resolution thermal infrared (TIR) images. *International Journal of Remote Sensing* 34: 4812–4831.

Whiting, P.J. and Bradley, J.B. 1993. A process-based classification system for headwater streams. *Earth Surface Processes and Landforms* 18: 603–612.

Wright, J.F., Moss, D., Armitage, P.D. and Furse, M.T. 1984. A preliminary classification of running water sites in Great Britain based on macro-invertebrate species and prediction of community type using environmental data. *Freshwater Biology* 14: 221–256.

Zachman, W.R. 1984. A river classification system: management for Minnesota's rivers. In: *Proceedings of the 1984 River Recreation Symposium, October 31–November 3,* Louisiana State University, Baton Rouge, LA, pp. 619–627.

Zernitz, E.R. 1932. Drainage patterns and their significance. *Journal of Geology* 40: 498–521.

Zumstein, J.F. and Goetghebeur, P. 1994. *Typologie des Rivières du Bassin Rhin-Meuse*, Rozérieulles: Agence de l'Eau Rhin-Meuse, 6pp.

CHAPTER 8

Modelling catchment processes

Peter W. Downs[1] and Rafael Real de Asua[2]

[1]Plymouth University, Plymouth, UK

[2]Stillwater Sciences, Berkeley, CA, USA

8.1 Introduction

Modelling of catchment processes and fluvial system response is dependent upon understanding a variety of fundamental forces that act as 'drivers' of the individual processes. Over time-scales relevant to most geomorphological studies ($\sim 10^1$–10^4 years), the majority of the drivers relate to independent catchment-scale variables including both natural phenomena and responses to human activity. Such variables have long been argued (Schumm and Lichty 1965) to provide the basis for the shorter-term ($\sim 10^0$–10^2 years) interdependent relationship between fluxes of flow and sediment and the channel morphology. Understanding the role(s) and hierarchical scaling of these numerous variables over extended periods provides the intellectual justification for developing catchment-based process models but also hints at the inherent complexities involved in achieving a detailed and robust output, especially given the complex relationship between initial topography, processes and eventual equilibrium end-points (Lane 1998; Perron and Fagherazzi 2012).

So fundamentally intertwined are the issues inherent to both catchment process modelling and fluvial geomorphology that teaching in fluvial geomorphology frequently begins with a conceptual model of catchment processes – notable examples include fig. 1 in Knighton (1998, p. 2) and fig. 7.2 in Thorne (1997). Working upscale, the dynamics of process and form at the scale of the individual channel cross-section can be understood through analysis of the channel form and its geotechnical properties (see Chapter 11), but such changes are controlled by interactions between flow hydraulics, the channel planform, the sediment composition of the channel bed and banks and local in-stream and riparian vegetation in the surrounding river reach (see Chapter 19). Many of these factors are in turn controlled by the dynamics of flow and sediment over longer reaches (e.g. stretches between major tributaries) and can be approximated through modelling flow hydraulics and sediment transport (see Chapter 18). However, the boundary conditions for these two parameters are determined by the channel network hydrology and sediment transfer processes (see Chapter 16), which combine to determine the frequency and distribution of geomorphologically effective flow events and the dynamics of sediment supply, transport and deposition. Ultimately, network hydrology and sediment transfers are themselves functions of the hillslope hydrology and sediment yields from terrestrial surfaces within the catchment. Therefore, the fundamental process drivers of fluvial geomorphology become factors such as the contemporary regional climate, topography and land cover set within the context of recent weather events and allied to a set of individual system history features that define the 'uniqueness' (Schumm 1991) of the individual catchment (Haff 1996). Classically, these may include the catchment's climate history since the beginning of the Holocene period or longer and contemporary studies are generally remiss if they do not account for the ever increasing influence of human activities. The dominant suite of geomorphological processes involved in any individual (sub-)catchment ultimately depends upon the catchment geology and lithology set by its 'geomorphic province' of physiography, bedrock type and structure and climate history (Montgomery 1999).

Geomorphologists intent on modelling fluvial geomorphological processes and the dynamics of resulting river systems therefore have a significant challenge in integrating these various influences over relevant time frames. Unsurprisingly, many types of fluvial geomorphological models have developed at a number of different scales, not all of which are appropriate to catchment-scale application (Van De Wiel *et al.* 2011). In this chapter, we focus on model types with an avowedly catchment-scale potential for application: various other topics in modelling in fluvial geomorphology are dealt with elsewhere in this volume (see Chapters 17–21) and in periodic reviews (e.g. Merritt *et al.* 2003; Martin and Church 2004; Codilean *et al.* 2006; Tucker and Hancock 2010; Odoni and Lane 2011). We begin by outlining the relative merits of several approaches available for developing a catchment process model before examining the various approaches in more detail. We subsequently consider tools for developing a catchment process model focusing on issues of representation and accuracy and the prospects for future model developments.

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

8.2 Approaches to catchment processes modelling

There are numerous classifications of model types in geomorphology (at least 10 in a recent review; Odoni and Lane 2011) and many of the alternatives are represented in catchment process modelling. To begin, there is a spectrum of catchment model possibilities ranging from conceptual models to those that are fully numerical. Assuming, as did Baker (1996), that geomorphology relies primarily on 'retroductive inference', wherein the instinctive reasoning of geomorphologists leads them to move from real effects (based on observations) to real causes (deduction of processes that must link cause and effect), then catchment conceptual models are an inherent first step in process modelling. Such logic explains why the conceptual model of Knighton (1998) provides an underpinning for teaching about process and form in fluvial geomorphology and why that of Dietrich and Dunne (1978) has provided the inspiration for many catchment sediment budgets (see Chapter 16, Fig. 16.4).

Developing from such generic conceptualizations are 'interpretative' approaches to catchment process modelling that are not fully quantified, but which use a combination of data sources (ranging from empirical to narrative) and expert analysis and observation to construct a 'semi-quantified' catchment process model that overcomes many of the spatial and temporal scale problems common to fully numerical forms of catchment modelling. Such models are often designed with the intent of problem solving for applications in river management and restoration and the ability to incorporate influential historical data is an important attribute that has always proved highly problematic in more numerical approaches. Examples include the 'Fluvial Audit' in the United Kingdom (Sear *et al.* 1995, 2009) and the 'River Styles®' framework in Australia (Brierley and Fryirs 2005), whereas in the United States, the 'watershed analysis' approach (Montgomery *et al.* 1995) may use a combination of catchment models within its overall structure, potentially including overlays in GIS intended to provide exploratory understanding related to the 'process domain' concept (Montgomery 1999).

An alternative to the use of expert judgment and interpretation is to develop fully quantified catchment process models. A first category includes models driven primarily by empirical data and that can be labelled 'data-theoretic' rather than 'model-theoretic' (in which model theory is the primary driver for model development; Odoni and Lane 2011). Indeed, the historically most popular catchment model in fluvial geomorphology is the data-theoretic sediment budget, vaunted as potentially providing a unifying framework for studies of fluvial geomorphology (Slaymaker 2003, 2008) and warranting a dedicated chapter in this volume (see Chapter 16). Sediment budgets provide an empirical mass balance of sediment transfer processes and rely heavily on an accurate conceptual model of catchment processes over which data are laid: differences in annual mass fluxes between budget components can be interpreted to indicate the most dominant processes, changes in catchment sediment dynamics and in-channel morphology. An approach using process equations to simulate the sediment budget components (Prosser *et al.* 2001) moves the approach towards numerical modelling.

'Model-theoretic' approaches to catchment process modelling now include several categories of numerical simulation model. Unlike field-based empirical models, these models require a series of pre-assigned rules governing the process feedbacks between cells in a gridded or otherwise zoned digital elevation model of the catchment. Developing from distributed hydrological models, one category of these numerical models is process based on an 'explicit numerical reductionism' (Murray 2007). Such 'erosion and sediment transport' models (Merritt *et al.* 2003) are highly complex, are highly data intensive and focus primarily on fluxes operating on a largely static terrain surface. Almost in response, a second category of models in fluvial geomorphology has developed as a sub-set of 'landscape evolution models' (see Martin and Church 2004; Codilean *et al.* 2006; Tucker and Hancock 2010). There are various approaches available to represent the transport laws implicit to landscape evolution models (Dietrich *et al.* 2003) and in the 'reduced complexity modelling' (RCM) approach (see Brasington and Richards 2007; Murray 2007), the accuracy of process representation is forsaken to permit pseudo-realistic change in fluvial landforms over geomorphological time-scales. RCM approaches are therefore perhaps best considered as 'effect of process based' rather than 'process based'. In general, numerical models of landform evolution offer the enticing prospect of 'what-if' scenario modelling of future conditions, but face considerable challenges of computing power when applied over extensive catchment areas.

Because catchment processes are subject to complex scaling arrangements and include a significant historical component, there is generally a basic trade-off between breadth versus depth of model so that the choice of an appropriate catchment model will depend partly on its purpose. Ultimately, a decision about which catchment model is the 'best' will depend upon the objectives of the modeller. For instance, in scientific exploration, the concept of 'emergence' in modelling is important (Kirkby 1996) and, in this regard, the potential for RCMs to help establish a better understanding between geomorphology form and process is high. However, where the emphasis is on spatially explicit prediction, reductionist modelling approaches may be more appropriate unless their data requirements cannot be met, in which case an empirical or interpretative model might be more practicable. We return to this theme in the final Prospect section of this chapter.

8.3 Conceptual models

Catchment process modelling originated with late nineteenth century conceptual 'word–picture' models of landscape evolution (Tucker and Hancock 2010), most notably in the geographical cycle of William Morris Davis (1899), but also those of Penck (1953) and King (1962). These models provided an initial attempt at *explaining* how landscapes might evolve at large scale, using space-for-time observations of morphological change to depict evolution and conjecture as the basis for process understanding (Martin and Church 2004). Despite the mid-century 'revolution' in geomorphology towards reductionist investigations of landscape processes (Gregory 2000, pp. 63–66) and the development of 'quasi-mechanistic' hillslope process models of Kirkby (1971) and Ahnert (1976) (Martin and Church 2004), word–picture conceptual models were still the only practicable way of representing landscape evolution models until the 1980s heralded the beginnings of the digital era (Tucker and Hancock 2010). Indeed, as noted, more recent conceptual models that make an explicit connection between catchment processes and resultant morphology (Dietrich and Dunne 1978; Thorne 1997; Knighton 1998) are still useful in the initial understanding of catchment processes and are the fundamental basis of 'more sophisticated' catchment models. Conceptualization itself is argued to represent the formal process of encoding a perceptual model of the mind (or group of minds) with a series of properties likely to include objects, processes, boundaries, boundary conditions and exogenous drivers that cause change in the boundary conditions (Odoni and Lane 2011). Such properties become the building blocks of process modelling.

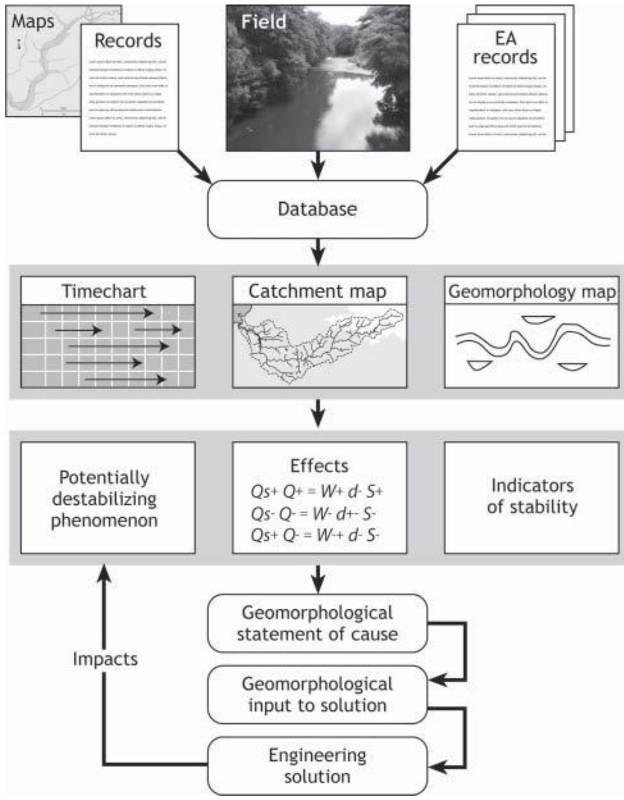


Figure 8.1 Procedure for undertaking a fluvial audit, using expert understanding to guide the interpretation of effects and the identification of potentially destabilizing phenomenon (PDP). Source: Newson and Brookes, 1995. Reproduced with permission of Wiley.

8.4 Problem-centred interpretative models

Interpretative models develop from a conceptual framework by integrating observations of catchment conditions using expert analysis to overcome the spatial and temporal scale problems common to other more data intensive forms of catchment modelling. Data collection is frequently based on field reconnaissance to provide structured field observations via a rapid assessment protocol (pioneered by Kellerhals *et al.* 1976). With explicit guidelines for field interpretation (see Downs and Thorne 1996; Thorne 1998), replicable process observations can be achieved. The resulting observations are often paired with readily-available digital data sets providing catchment coverage: sources such as aerial photographs, historical maps and data related to catchment terrain, land cover and geology can be integrated via a GIS to provide a contextualized understanding of prevailing and historical catchment processes.

Many interpretative models have been developed world-wide in response to practical concerns in river management (see Downs and Gregory 2004). The 'Fluvial Audit' (Sear *et al.* 1995, 2009), for example, develops a partially quantified appraisal of field and archival data relating to sediment generation, transport and deposition across a catchment. Combined with an implicit conceptual understanding of the processes linking the sediment components, catchment and reach-scale geomorphological maps are developed alongside a chronology of impacts with potential implications for the reach of management interest. From these outputs, the most likely causes of the management problem are identified as the basis for a cause-effect approach to sustainable river channel management (Fig. 8.1). The approach, applied to over 50 catchments in the United Kingdom by 2003 (Sear *et al.* 2003), has since been extended to use spatial data within a GIS to provide a weighted, multi-criteria assessment of the degree of naturalness and modification of the river reach (Sear *et al.* 2009) as the basis for decision-making. Using similar field and data sources, the River Styles® framework (Brierley and Fryirs 2005) has been applied widely in river management applications in Australia. The approach consists of a baseline survey as the basis for a catchment-framed assessment of river condition including the historical evolution of the channel. These stages are the basis for interpreting likely future changes and recovery potential as part of a catchment-based vision to guide management applications.

'Watershed analysis models' (Reid and McCammon 1993; Montgomery *et al.* 1995) also represent an interpretative approach to catchment modelling that responds to needs for a cause–effect, historically informed approach to catchment management. The basic goal of watershed analysis is '… to generate a spatially explicit understanding of a landscape and its ecosystems at a resolution sufficient to allow assessment of the integrated environmental consequences of inherently local land use practices' (Montgomery *et al*. 1998, p. 244). Montgomery *et al*. (1995) argue that the fundamental basis of watershed analysis is fourfold (Fig. 8.2). First, an initial stratification of the catchment is made to subdivide 'landscape-level' functions and processes according to dominant processes. Archive data are then collected in order to reconstruct historical conditions prior to collecting data and observations related to contemporary conditions, a phase that will include field analysis. Finally, following the application of physically based models of the dominant processes in their historical context, the landscape is re-classified according to the management questions. Like the fluvial audit and River Styles® approaches, watershed analyses provide an effective form of exploratory analysis focussed on prompting catchment-specific questions for more detailed analysis, rather than a device for deriving generic process insights.

Because each watershed analysis will vary in detail according to the level of confidence required to determine acceptable risk in the outcome (Montgomery *et al.* 1995) and because the process is critically dependent on the initial stratification to determine dominant processes, the models suitable for analysis will vary between applications. In catchments with well-defined breaks in geomorphological processes and similar land uses, such as the steep, forested, catchments of the Pacific Northwest of the United States, Montgomery *et al.* (1998) identify landslide-derived sediment generation predicted from a physically based model of potential slope instability (SHALSTAB, Montgomery and Dietrich 1994). SHALSTAB incorporates topographic controls, soil type, hydrology and rainfall data developed from digital terrain data to offset the incomplete record obtained from aerial photographs and maps, and a terrain-derived slope threshold criteria to define the headwater extent of the channel network (Montgomery and Foufoula-Georgiou 1993). Reid *et al*. (2007) used SHALSTAB in conjunction with the hydrological model TOPMODEL (Beven and Kirkby 1979) to examine the connectivity of landslide-generated sediment delivery to channel networks. A related modelling approach is to create a classified overlay of data such as geology, land cover and hillslope gradient data in a GIS to provide the initial stratification of coherent process domains (Montgomery 1999) of sediment source and transfer processes. The approach, used in conjunction with field validation, should produce a spatial differentiation of similar unit-area sediment production rates (Booth *et al.* 2014). The method can form a contributory part of the sediment budget process (see Reid and Dunne 1996) or can be used directly to inform process-based management: applications in steep catchments, include identifying areas with potential for high production of coarse or fine sediment which may be particularly significant in assessing fish habitat (Downs and Booth 2011).

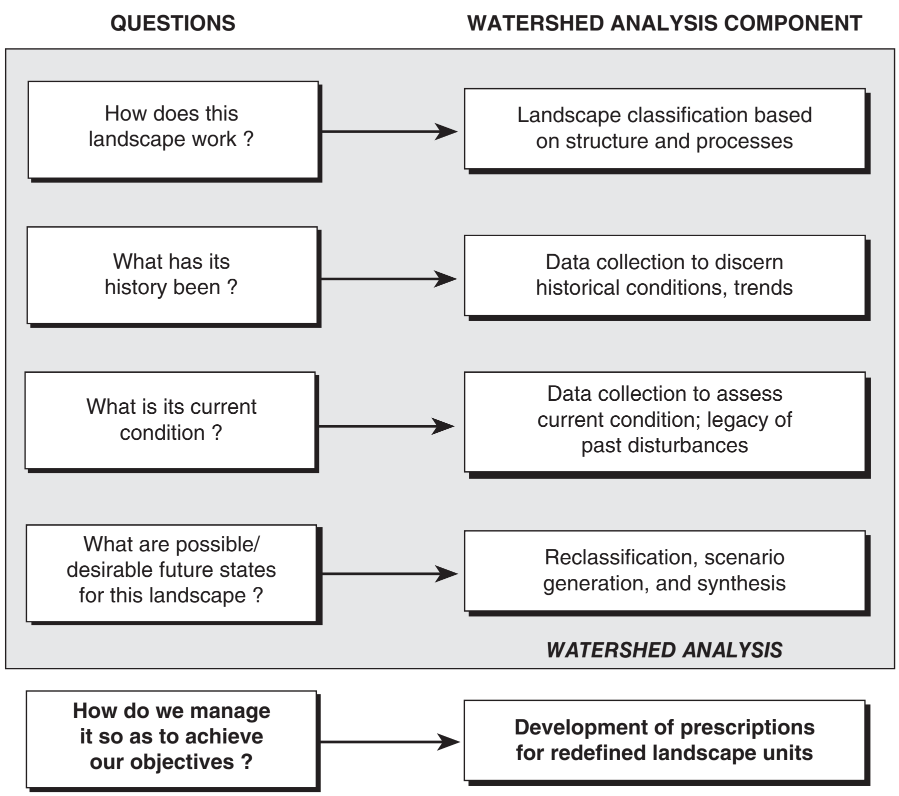


Figure 8.2 Overview of the watershed analysis procedure, illustrating major questions asked of the modeller at each stage. Source: Montgomery *et al*, 1995. Reproduced with permission of Wiley.

Conversely, in watershed analyses in lowland England, where there is no direct coupling of hillslope and channel processes and a greater heterogeneity of land use types and histories, the initial stratification of dominant process regimes in the context of human impacts can be the central question (Downs and Priestnall 1999).

Approaches using interpretative modelling therefore require a conceptual model of catchment process (generally used to frame the problem statement), a system of organized field data collection and computer-collated catchment morphological data. Process insights are derived largely by association and, as evidenced above, these approaches are frequently used in problem-led rather than research-led applications. Used with care, they are a viable and practicable model technology where expert judgement can override problems of scale and data accuracy faced in direct numerical modelling, facilitating conclusions related directly to the observed adjustment of the fluvial system in its catchment historical context. They also provide considerable 'richness' (Kirkby 1996) in terms of the understanding achieved relative to the data input requirements. Interpretative approaches have developed alongside catchment-based concerns for sustainable environmental management and are often closely associated with cumulative watershed effects (Reid 1993) and estimation of catchment sediment budgets (Reid and Dunne 1996). Although they offer less potential for scientific insight into the fundamentals of catchment processes than do numerical models, they are suitable for development by small teams of researchers to explore locationally specific issues. Overall, because landscape-level stratification is not constrained by a pre-determined scale, interpretative models will need to be extremely flexible in order to be updated and revised as new material becomes available (Montgomery *et al.* 1995), hence the logical integration of digital data based on GIS as the analytical platform for development.

8.5 Data-driven empirical models

Another category of model that develops explicitly from a conceptual foundation is that of enumerated models of processes of catchment production, transfer and delivery, generally known as a sediment budget: '… an accounting of the sources and disposition of sediment as it travels from its point of origin to its eventual exit from the drainage basin' (Reid and Dunne 1996, p. 3). As detailed in Chapter 16, sediment budgets provide an empirical mass balance of sediment inputs, outputs and changes in storage that should account for patterns of sediment production, storage, transfer and rates of movement through storage across the catchment extent (Dietrich *et al.* 1982). As such, sediment budgets are argued to represent a powerful model of drainage basin sediment systems (Wasson 2002; Warburton 2011), raising the prospect that they provide a unifying framework for studies in geomorphology (Slaymaker 2003, 2008).

Sediment budgets have been constructed at several levels of detail, ranging from those based on spatially extensive, lumped estimates of sediment yield to those that are more directly spatially distributed and process based. The latter, in particular, allow the prospect both of documenting the impact of changing catchment conditions on fluvial system dynamics (Trimble 1999, 2009; Gregory *et al.* 2008; see Fig. 8.3), including channel morphological response, and deriving process insights from the comparison of mass balances. Examples are provided in Chapter 16. Sediment budgets collated directly from field data representing each of the processes important to that geomorphic province (Montgomery 1999) have the practical advantage in modelling terms of being developed

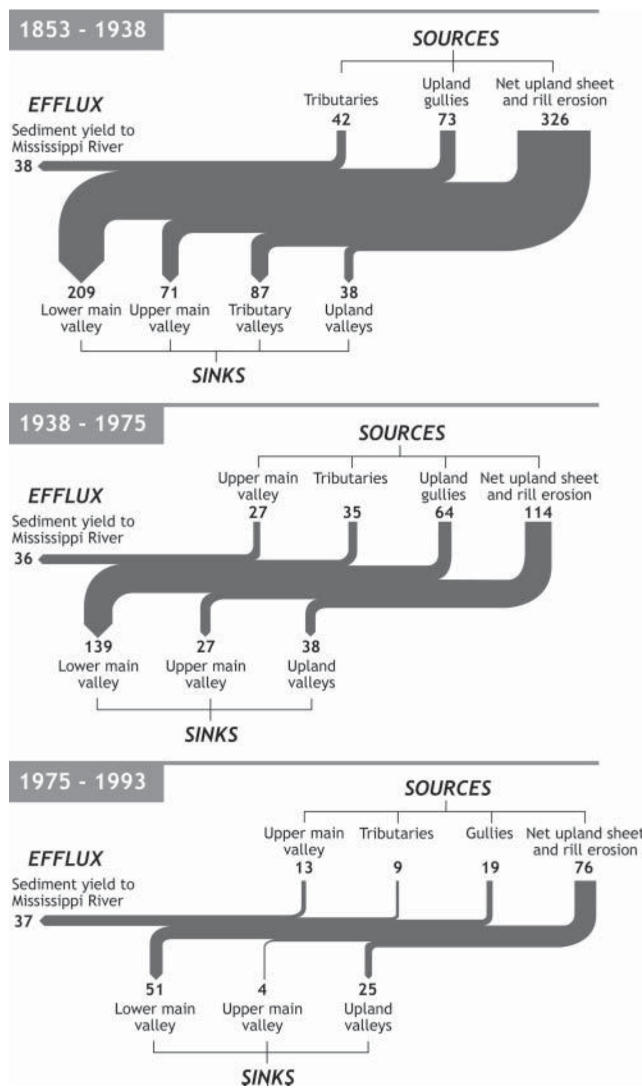


Figure 8.3 The use of a sediment budget model to plot changing fluvial systems dynamics over time in Coon Creek, Wisconsin. Note that catchment outputs are estimated to remain largely constant over time despite large variations in the input sediment sources. Source: Trimble, 2009. Reproduced with permission of Elsevier.

directly from a conceptual model of catchment functioning without the need for extensive rules regarding boundary conditions under which the model operates – indeed, part of the richness of a sediment budget comes from 'real-life' inferences that can be drawn from the enumerated processes. Such data-theoretic (Odoni and Lane 2011) inferences represent a rather inductive learning process and there are concerns that sediment budgets can confer an unwarranted and misleading impression of accuracy, especially when the mass balance is closed using unmeasured residuals (Kondolf and Matthews 1991). However, sediment budgets can be a very valuable tool in understanding the geography-specific process operation of a catchment in contrast to the more deductive but generally more abstract model-theoretic results achieved from numerical modelling (see the next section). Perhaps as a consequence of these attributes, sediment budgets are used both in scientific and practical applications.

The single largest problem in developing process-based, spatially distributed sediment budgets is that their data demands are rarely met by the availability of suitable monitoring data. An alternative to lumping the data over broad spatial extents is to use data layers to define catchment segments that are coherent in terms of their potential for sediment source and transfer (as described in Section 8.4, Problem-centred interpretative models) and then to extrapolate a sample of quantified field survey over the catchment extent. Changing volumes and sources of sediment can be accommodated using repeat aerial photographs (Reid and Dunne 1996). Data problems are exacerbated where intensive field surveys are precluded, for instance, where multiple sediment budgets are required for management purposes. An alternative scenario is to create a budget based primarily on sub-catchment data equations to simulate process-based estimates, rather than using empirical data. SedNet was developed in response to needs for extensive sediment assessments for land and water management issues in Australia (Prosser *et al.* 2001). Centered on a digital elevation model (DEM)-based river network with an area-based threshold for sediment supply and river links defined at nodes in the network, the model uses a variety of equations, often empirically calibrated, to develop separate budgets for bedload (Prosser *et al.* 2001) and/or suspended sediment (Wilkinson *et al.* 2009) (Fig. 8.4). Despite its attempt at a simple model structure, the cumulative input requirements of SedNet can often only be satisfied based on literature values (Merritt *et al.* 2003) and, because of the generalizations inherent to the contributing equations, the model is recommended for application over large spatial extents (>3000 km^2; Wilkinson *et al.* 2004).

Through the introduction of equations from which process estimates are derived, sediment budget simulations start to blur the distinction between sediment budgets as a purely empirical model and as a fixed-terrain numerical model, as described below. However, the use of either equation-driven devices such as SedNet or GIS-based methods for the extrapolation of quantified field survey has the important benefit of allowing spatially distributed sediment budgets to extend beyond the few square kilometres in drainage area characteristic of most directly monitored sediment budgets. Neither approach alleviates concerns for the need for independent corroboration of budgets and requirements for uncertainty assessment, both of which have been historically very limited.

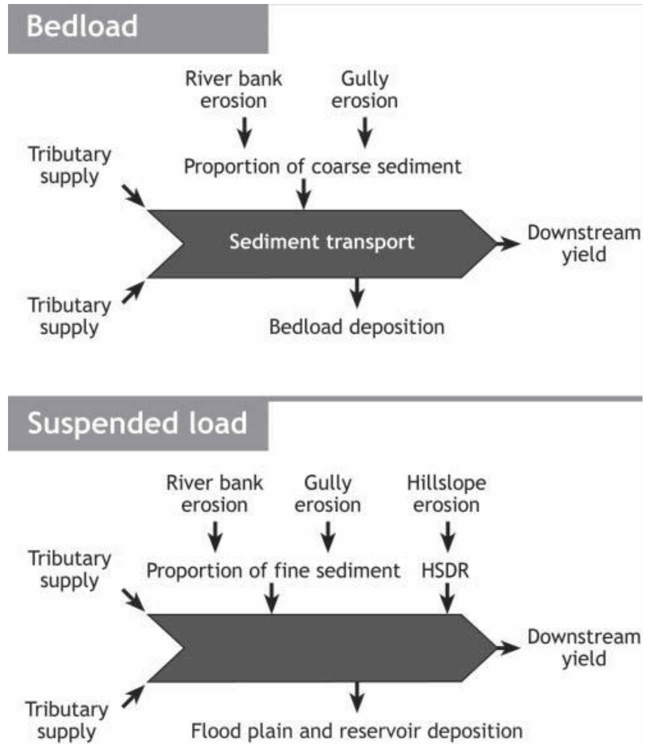


Figure 8.4 Schematic representation of the component entities estimated in developing either a bedload or suspended sediment budget using the SedNet model, Redrawn from Wilkinson *et al.* (2004, 2009).

8.6 Numerical models

In contrast to the direct empiricism of traditional sediment budgets, catchment-scale numerical models of fluvial geomorphology processes must accommodate the wide variety of hierarchical process drivers that control fluvial geomorphology (see Introduction). This implies an extensive suite of rules and boundary conditions capable of integrating a plethora of data sources measured over different spatial and temporal scales. Intriguingly, progress in over 25 years since a seminal review by Anderson (1988) demonstrates that models have been developed based on two opposing starting points (but see Prospect). To begin, there are bottom-up, reductionist models extended from numerical models in hillslope hydrology and soil erosion (e.g. WEPP, Flanagan *et al.* 1995; SHETRAN, Bathurst *et al.* 1996). Such physically based distributed models (PBDM, Birkinshaw *et al.* 2010) models are based on upscaling small-scale 'process correct' simulations and derive from a deterministic philosophy wherein perfect knowledge of the contributory processes would result in an invariant and repeatable fluvial geomorphological response. Sub-models of hillslope hydrology and soil erosion are coupled with hydrological and hydraulic routing models capable of simulating water discharge and sediment transport. These models, part of a larger collective of 'erosion and sediment transport' models (reviewed in Merritt *et al*. 2003), generate process fluxes over a fixed terrain and generally do not simulate the resulting evolution in catchment and river channel morphology. The development of a reductionist model is a long-term project requiring teams of researchers (see Flanagan *et al.* 2007) whose process sub-models will be at different stages of development or revision. Further, and as foreseen by Anderson and Sambles (1988), a significant investment of time is required by the user to understand and apply such complex models. Assembling the necessary distributed physical information to run a PBDM such as SHETRAN may take weeks of preparatory work ahead of the first test simulations (Birkinshaw *et al*. 2010), although basic simulations developed directly from GIS data using automatically derived river networks can be more rapid (Birkinshaw 2010).

As an alternative, the last decade has witnessed a surge in popularity of dynamic cellular model approaches to landscape evolution following influential developments in the 1990s (Willgoose *et al.* 1991; Murray and Paola 1994). These models are also physically-based but centre on a top-down 'synthetic' (Paola 2001) or 'holistic' (Van De Wiel *et al*. 2007) approach wherein rules of process representation (i.e. the governing physics) are 'relaxed' (Coulthard *et al*. 2007). Such so-called 'reduced complexity models' (RCMs) (see Brasington and Richards 2007) simplify the incorporated process components and focus instead on achieving a physically based accuracy in terms of morphological evolution of the river channel, river network or catchment terrain as the input variables are adjusted. This suite of models is less computationally demanding and is more amenable to experimental scientific application and development by individuals and small teams of researchers: they aim to simulate accurately the types and styles of adjustment rather than to provide locationally precise information about geomorphological change, but they do involve an explicit temporal component. A brief review of these model types follows.

Process-based, reductionist models

For reasons advanced in the Introduction, there are enormous challenges in achieving 'explicit numerical reductionism' (Murray 2007) in 'process correct' catchment-scale models of fluvial geomorphology. The Water Erosion Prediction Project (WEPP) model, for instance, required 10 years of development before its release as a comprehensive software package in 1995 (Flanagan *et al*. 1995, 2007). WEPP uses 'fundamentals of stochastic weather generation, infiltration theory, hydrology, soil physics, plant science, hydraulics and erosion mechanics' (Ascough *et al*. 1997, p. 921) to combine process models and physically based empirical relationships to simulate numerous aspects of the hydrological cycle and consequent soil erosion. Designed initially for application in agricultural field management (Foster and Lane 1987; Lane and Nearing 1989) as a process-based successor to the widely used, empirically based universal soil loss equation (Wischmeier and Smith 1978), WEPP is now used extensively to assess soil erosion and evaluate remediation efforts in public forests and rangelands in the United States and elsewhere (Flanagan *et al*. 2007). An alternative, SHETRAN (Bathurst *et al.* 1996; Ewen *et al.* 2000), is a cell-based distributed model that extends the Système Hydrologique Européen (SHE) distributed hydrological model (Abbott *et al.* 1986a,1986b) to include algorithms for the generation and transport of sediment (including those of soil erosion, based on a sub-model, SHESED; Wicks and Bathurst 1996), 'modelled either by finite difference representations of the partial differential equations of mass and energy conservation or by empirical equations derived from independent experimental research' (Bathurst *et al*. 1996, p. 356). Intended primarily for applications related to surface water and groundwater resources management, each physical variable in SHETRAN is represented by one parameter in each cell (Fig. 8.5). Geomorphology-based applications include those related to landslide assessment (e.g. Bathurst *et al.* 2005; Bovolo and Bathurst 2012), land use impacts on storm runoff and sediment yield (e.g. Bathurst *et al.* 2007) and relationships between catchment area and sediment yield (e.g. Birkinshaw and Bathurst 2006).

Spatial representation is fundamental to reductionist models because representing catchments of different sizes involves more than simply scaling up the fundamental units at which the model operates: as size changes, emphasis shifts from a concern for slope micro-topography and individual flow hydrographs at the hillslope scale, to topography, soil and vegetation patterns over longer periods at the catchment scale and to lithology and climate extending back over thousands of year at the regional and global scale (Kirkby *et al.* 1998). For catchment applications, WEPP is calculated initially on a field-by-field basis with routing through the catchment achieved by assigning a topological format to each slope linking distinct 'overland flow elements' to 'channel elements' and 'impoundment elements' and invoking hydraulic process models (Ascough *et al*. 1997; Williams *et al*. 2010) (Fig. 8.6). In SHETRAN, the model is usually run for a maximum of a 50×50 grid whereby the user can increase cell size according to the catchment size, but this capability operates effectively only when the cell size is small relative to variations in the local hydrological controls and responses (Bathurst *et al.* 1996). Once the grid cells are larger than the distance over which parameters show significant variation, then model accuracy is likely to decrease rapidly as output errors in one cell generate input errors in the next. To tackle this concern, spatially integrated 'effective' parameter calculations have been developed and applied with reasonable success for the 701 km^2 Cobres catchment in Portugal using a 2 km $\times$ 2 km grid cell (Bathurst *et al.* 1996), and an upper limit catchment size of 2500 km^2 is advised. Likewise, in WEPP,

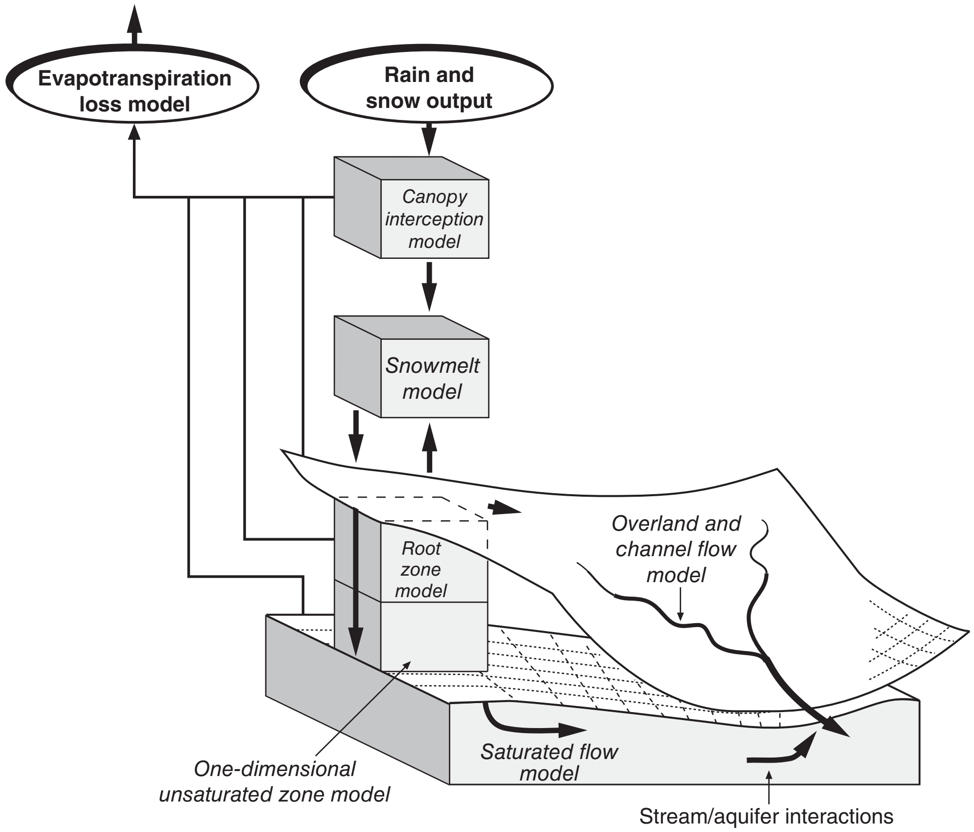


Figure 8.5 Schematic diagram of the major components forming the SHETRAN cell-based integrated component process model. Adapted from Dunn *et al*, 1996.

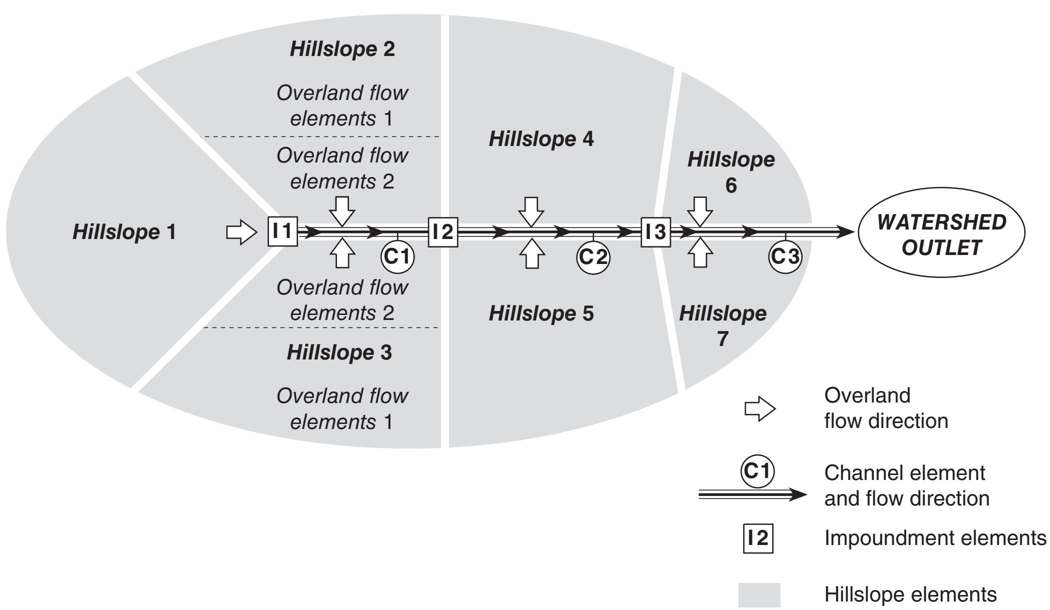


Figure 8.6 Schematic representation of the application of the WEPP model over a small watershed to include the integrated components of multiple hillslopes, channel processes and impoundment structures. Source: Ascough *et al*, 1997. Reproduced with permission of American Society of Agricultural and Biological Engineers.

sensitivity testing on progressively longer hillslopes indicated that the model becomes less reliable owing to changes in the dominant controls on processes (Baffaut *et al*. 1997). The recommended maximum hillslope length is 100 m, restricting the permissible size of an individual catchment to approximately 2.6 km^2. Such 'small watershed' applications of WEPP to estimate surface runoff and sediment yield include forested (Dun *et al*. 2009) and agricultural settings (Williams *et al*. 2009, 2010). The addition of a new channel routing model (Wang *et al*. 2009) to the software in 2012 offers the prospect of applying the model to somewhat larger catchments, but the lack of explicit scaling factor means that larger application areas will greatly increase data requirements, model complexity and the potential for error accumulation (Merritt *et al*. 2003).

One considerable constraint for the fluvial geomorphology application of reductionist models is that they do not allow for feedback from erosion and deposition processes to alter channel morphology (for instance, in allowing channel bar deposition to compensate for bank erosion). SHETRAN can be used to route channel sediments and predict bed armouring processes (Wicks and Bathurst 1996), but the spatial locations of erosion and deposition were not well predicted. Subsequently, Birkinshaw and Bathurst (2006) suggested that SHETRAN simulations are most relevant over time-scales of a decade or two (i.e. periods where there is unlikely to be significant change in channel morphology) and therefore can be used, for instance, to test the assumed inverse relationship between sediment yield and catchment area (Birkinshaw and Bathurst 2006). WEPP does incorporate some channel morphology components, but Ascough *et al*. (1997, p. 922) suggested that the model is suited to application in constructed waterways and concentrated flow in gullies, but not to perennial stream channels, locations with dynamic contributing areas, channels with mass bank failures or headcut erosion processes or where erosion is generated by seepage effects.

Effect-of-process-based reduced complexity models

In contrast to bottom-up reductionist approaches that upscale small-scale process simulations to represent catchment geomorphological processes, the 'reduced complexity' approach represents a physically based top-down approach wherein the primary modelling objective is to simulate the dynamic morphological behaviour of a landscape to the effects of (changes in) the dominant processes. The development and popularization of RCMs can be attributed, in part, to the problems faced by reductionist approaches in trying to represent dynamic landforms such as alluvial channels (Coulthard *et al.* 2007, and see above) and their introduction represents a potentially important contribution to fluvial geomorphology.

Models of fluvial landscape evolution, commonly labelled 'landscape evolution models' (LEMs) or 'surface process models' (SPMs) (Codilean *et al*. 2006), arose during the 1990s as increasing computing power permitted iteratively better modelling of the 'geomorphic transfer functions' that govern landscape evolution (Martin and Church 2004; Tucker and Hancock 2010). While various numerical methods are available to drive LEMs, the increasing availability of gridded DEMs meant that the cellular automaton method, in which spatial transfer functions are distributed via a series of simplified (physically based) rules and time is simulated via discrete updates across all cells (Fonstad 2006), has become popular as a means of modelling the dynamics of fluvial system evolution. A variety of LEM models have been established, including those focused initially on hillslope evolution (e.g. SIBERIA, Willgoose *et al*. 1991) or floodplain evolution (Murray and Paola 1994). Later models held the promise of catchment-scale evolution with a focus on the fluvial system (e.g. CAESAR, Coulthard *et al*. 1999). One advantage of the cellular models over the reductionist approaches is that their computation efficiency allows the possibility of modelling fluvial system evolution over large spatial extents (up to 1000 km^2) and over historic or Holocene time-scales (Nicholas 2005), allowing the potential of simulating system response to environmental change (Van De Wiel *et al*. 2011).

Cellular LEMs operate by simulating hydrological, fluvial and hillslope processes to route water and sediment over a surface represented by a catchment DEM (Van De Wiel *et al*. 2011). Usually, fluvial processes are routed along single or multiple paths represented by the steepest gradient/s between the cell and its immediate neighbours (see Murray 2007; Nicholas and Quine 2007; Wilson *et al*. 2007), although the rules can be modified for processes governed primarily by bed elevation, such as bedload transport, rather than by the combined elevation of the bed and water column depth, such as suspended sediment transport (see Van De Wiel *et al*. 2007). According to a suite of rules assigned for simulating processes of water flow, sediment erosion, transport and deposition and morphological change, the elevation of individual grid cells is altered, allowing the host DEM to evolve ahead of the next iteration of the model. Thus, over the course of thousands of iterations, the landscape and fluvial system evolve according to feedback between slope and fluvial processes. Of course, the notion of a reduced complexity model of cellular landscape evolution is somewhat misleading, partly because all models are, by definition, simplifications of reality (Brasington and Richards 2007), and partly because although the governing rules may be simple, the emerging model outputs may be non-linear and complex. For example, the routing of water over the DEM surface is dependent on the shape of the grid, model rules governing lateral movement of water in areas of shallow terrain and whether flows become over-concentrated into topographic low points (Nicholas 2005). Also, in the simulation of reach-scale fluvial dynamics into an RCM, the requirement to simulate lateral erosion processes is conceptually challenging as it does not follow directly from other flow routing and sediment transport processes (Van De Wiel *et al*. 2007).

Even RCMs struggle with the scaling and data requirements of catchment-scale modelling. In catchment applications of the

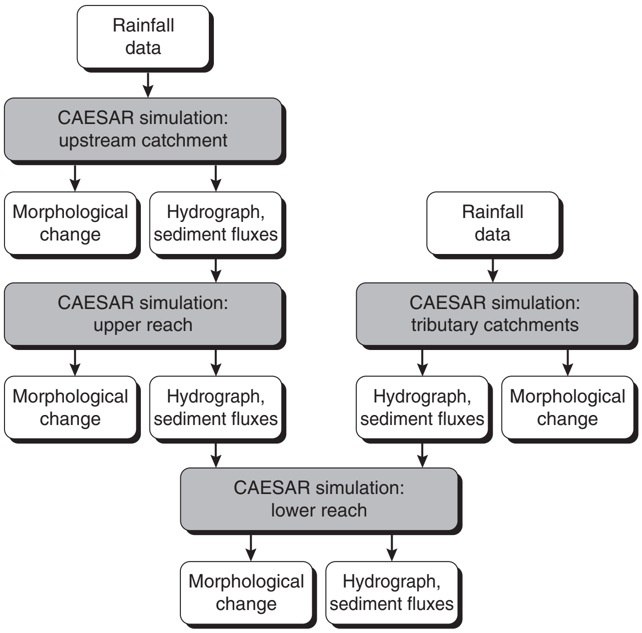


Figure 8.7 Schematic representation of catchment-scale application of the CAESAR model to the upper River Severn in Wales, using catchment-scale inputs to derive the boundary conditions for a more detailed reach-scale simulation at the reach scale near the town of Caersws. Source: Van de Wiel, 2007. Reproduced with permission of Elsevier.

CAESAR model (Coulthard *et al.* 2000, 2002), a hierarchical approach is utilized with the catchment model driven by the DEM and hourly rainfall records providing inputs to a more detailed model of fluvial dynamics in the reach of interest (Fig. 8.7). There are clear parallels here with some of the scaling approaches required for reductionist models (see above) and, indeed, with investigations that set the 'catchment context' for interpretative models. The hierarchical approach efficiently derives the boundary conditions for the reach of interest in the probable absence of conveniently located monitoring stations and allows for computational efficiency by using a coarser scale for the sub-catchment model components and provides opportunities for parallel computing (Van De Wiel *et al.* 2007). Likewise, in modelling the impact of environmental change, LEMs have to reconcile the various spatial and temporal scales of process operation within the model, including the considerable issue of determining appropriate initial conditions for the model (Van De Wiel *et al.* 2011).

8.7 Tools for developing a catchment process model: representation and accuracy considerations

Capabilities for catchment process modelling have developed since the 1980s due to the advent of digital data sets, better process understanding of 'geomorphic transfer functions' (Tucker and Hancock 2010), GIS capabilities for terrain modelling (Moore *et al.* 1994), popularization of the cellular automata approach (Fonstad 2006) and ever-increasing computing power, to name but a few technological and intellectual advances. Since the first edition of this chapter (drafted in 2001; Downs and Priestnall 2003), numerous advances have occurred, including that the resolution (and availability) of catchment terrain and thematic data layers now rarely constrain catchment modelling; high-resolution airborne light detection and ranging (LiDAR) data are now the readily available foundation of many catchment models (and amenable to regular, customized updates at relatively low cost); reductionist approaches to numerical modelling of catchment process have been challenged in popularity by RCMs stemming from cellular landform evolution modelling; and improved GIS functionality makes exploratory modelling of catchment data more viable than ever before. Two themes critical to the provision of tools for catchment process modelling include the representation of catchment data and issues related to the performance of the resulting models.

Input data representation

Modelling catchment-scale influences on fluvial geomorphology is inherently data demanding. It is very important to choose a digital representation structure that is both sufficient for the required scale and resolution of the modelling but avoids consuming unnecessary computational resources. Nationally standardized DEMs and high-resolution LiDAR data are increasingly used over catchment extents, and very high-resolution floodplain and channel surveys can now be achieved cheaply via remotely controlled sensors set on small drone aircraft or helium balloons (Moorhead *et al.* 2012; Wallace *et al.* 2012; Glennie *et al.* 2013) if not by ground-based LiDAR. While Brooks and Anderson (1998) rightly warn that a search for ever improving model resolution can be a dangerous distraction from improving the process basis of such models, concerns for input quality data to catchment models should figure highly in the geomorphological modeller's mind, particularly given the impact of terrain representation on geomorphological processes (Lane *et al.* 1998a). There is an inherent risk that uncertainties in digital data representation will mask uncertainties in the fluvial geomorphological process modelling and could be propagated throughout the entire analysis. Understanding and management of the uncertainty in data, the importance of data registration in multi-year evolution analyses and the use of metadata are vital if researchers are to be able to make full and appropriate use of digital data sets. Where digital data sets are created manually from surveyed data (also now achieved at increasingly high resolution using differential and 'map-grade' GPS and robotic total stations), uncertainties related to the sampling strategy, data abstraction from raw field measurements, choices regarding storage method, types of algorithm used to derive new parameters and suitable levels of data resolution and interpolation will, to a large extent, determine the errors inherent to the model building process (see Lane *et al.* 1998b). Several central

Table 8.1 Best-practice attributes for catchment modelling of geomorphological processes within a GIS.

Attribute	Comment
Core functions	
Capture and storage	Important to provide easy importation and integration of data sets to a single platform. Most GIS can now readily integrate field-sampled point data and, increasingly, systems facilitate the integration of data captured using LiDAR, ADCPs, global positioning systems (GPS), data loggers and other surveying tools
Organization and retrieval	Organize data sets for easy retrieval and display of their information either individually or in combination with other data sets. Multiple data set overlays are facilitated by GIS geo-referencing to 'real-world' spatial coordinates and this should also facilitate the management of attribute information associated with features or whole data sets
Data modification	Allow easy data editing to keep data sets up-to-date. Most GIS include tools for editing raster and vector data, including the ability to change or supplement the spatial information and attributes associated with the graphical features. Certain analytical operations in GIS may require data to be structured in such as way that the spatial relationship between features is stored, in addition to data themselves
Data exploration	
Derived parameter estimation	Provide tools to manipulate and analyse spatial information over the catchment extent, e.g. the use of distance functions such as *buffering* to derive zones of influence around river channel networks. Use *matrix algebra* to produce derived parameters, overlaying numerous raster-based representations according to mathematical criteria to produce previously unknown spatial information
Terrain indices and associated functions	Generate terrain data to provide exploratory relief maps and three-dimensional views of the catchment and the parameters distributed over that catchment. Terrain data may now be analysed using several topographic and hydrological functions commonly available in 'generic' GISs such as Idrisi, GRASS or ArcGIS (see Table 8.3)
Visualization	Allow the desired layer to be displayed over a two- or three-dimensional catchment visualization and from different perspectives. Allowing other feature attributes to be draped over surfaces, including the results of analytical queries based that may not normally be in the visible domain such as error or uncertainty
Model implementation	
Coupling of models and GIS	Consider model complexity: there is usually an inverse relation between model complexity and the ability to create that model directly in GIS, due to the constraints imposed by the GIS (Goodchild *et al.* 1996). It is often advisable to maintain only a loose coupling between complex models and the GIS
Custom functions	The flexibility of GIS is being improved through custom functions written in a conventional programming language rather than macros of the host GIS (e.g. C++ custom functions for the GRASS GIS; Batelaan *et al.* 1996). Increasingly custom functions are being written in the Python programming language, but the developer should be wary of making the system more complex and difficult to maintain than one based upon a set of generic GIS operations

concerns are introduced below; some best practice attributes for catchment modelling via a GIS are introduced in Table 8.1.

Surface topography

DEMs offer a representation of Earth surface elevations in digital form, usually as a regular grid of elevation values and the density and distribution of surveyed height information is fundamental in ensuring control over the resulting DEM (McCullagh 1998). Data sources for DEMs can include digital contours (Tyler and Greenlee 2012), stereographic or digital photogrammetry from aerial surveys, radar-based methods such as interferometric synthetic aperture radar (IFSAR) and LiDAR. Airborne LiDAR has become the standard in large-scale elevation mapping since it provides high-density elevation data to decimetre or even centimetre accuracy and it is free from many of the problems associated with digital photogrammetry (Smith and Smith 1996; Smith *et al*. 1997). In LiDAR, the surface elevations of all features on the ground are captured and these surface objects such as buildings and trees can be separated from the ground, leaving a DEM (Jaafar *et al*. 1999). Improved hardware and storage capabilities are being used to investigate the use of the full light spectrum instead of a limited number of returns (Reitbeger *et al*. 2009; Rosette *et al*. 2011). Further, ground-based ('terrestrial') LiDAR is increasingly being used to complement catchment-scale airborne LiDAR, providing detailed information for channel morphology (including inclined surface of channel banks, bed texture and channel vegetation) at the scale of the river reach (Heritage and Hetherington 2007), and providing important detail in areas where airborne LiDAR provides only a low density of ground points, such as in steep areas or those with very heavy canopy. Other recent developments show promise in the use of LiDAR for obtaining bathymetric depths (McKean *et al*. 2009) and for using the LiDAR's full waveform to obtain vegetation metrics (Anderson *et al*. 2008).

In developing terrain elevation surfaces from DEMs, source elevation points are generally converted using different interpolation algorithms such as kriging, splines, inverse distance weighting (IDW) or, more commonly owing to its speed and processing efficiency, triangular irregular networks (TINs) (Burrough and McDonnell 1998). TINs are often eventually converted into raster datasets, because the computational simplicity of a regular grid makes it a more popular choice of surface representation (Burrough and McDonnell 1998). The digital terrain model (DTM) is the fundamental basis for deriving many of the components of catchment process models (see Table 8.2). The effect of interpolation errors on the accuracy of the model and on derived parameters such as slope, aspect and flow direction should be noted (Wise 1998). Wherever possible,

Table 8.2 Summary of fundamental components of catchment process modelling that are derived from digital terrain models (DTMs).

Operation	Description
Relief (including 3D visualization)	Digital representations of the terrain surface in a variety of display formats to explore the nature of the catchment. Colouring by height and hillshading may be particularly effective in highlighting subtle undulations in the terrain (hillshading especially). Three-dimensional views from any angle or altitude can be produced, either colouring the terrain by height or draping an image over the surface. Image drapes can include aerial photographs, remotely sensed images or gridded representations of any of the data sets for the catchment, including the derived parameters described below
Slope	A variety of algorithms allow the calculation of maximum slope at any cell in the terrain model. Slope may be an important constituent of indices representing geomorphological dynamics
Sinks/depressions	Depressions or sinks in the terrain model often result from errors in input survey data or through the triangulation process. These can be automatically identified and filled to prevent them being treated as areas of internal drainage
Flow direction	To model water flow through the catchment, the direction of flow from each cell can be calculated, resulting in a 'flow matrix'. In many flow matrices, several possible flow directions are possible from each cell, so unique directional 'values' are required so as to result in a realistic flow matrix
Catchment area upstream	From the flow matrix, cells contributing flow to any given 'outlet' cell (representing a point on a river, for example) can be automatically calculated. The resulting cells can form an 'area of interest' when overlaying other data sets to explore catchment scale influences (Downs and Priestnall 1999). With programming or customization, other spatial units of study such as flow strips on valley sides can be defined from the flow matrix
Network extent of valley/floodplain corridor	From the DEM, eventually combined with the channel network layer, possibility to determine floodplain features (Alber and Piégay 2011; Notebaert and Piégay 2013)
Flow accumulation	For each cell, the number of other cells that contribute flow to that point can be calculated and allocated to a new grid. One of the main uses of this function is to define a channel network along cells of high flow accumulation and thence to use this network in routing operations. The channel network is defined according to cells exceeding a threshold value of flow accumulation. Conversely, cells having very low flow accumulation values can be taken to represent ridges or watersheds
Aspect	Aspect, derived from the direction of steepest slope at each cell, can be used, for example, to estimate the effectiveness of sunlight in increasing the rate of evaporation or snowmelt
Curvature	The terrain curvature in both plan and profile about a cell can be calculated and can contribute to calculations of hillslope hydrology parameters such as flow convergence
Flow length	Computes the distance from a cell downstream to the outlet or upstream to the divide
Stream order	Calculates Strahler and Shreve stream orders.

geomorphological 'objects' represented by sharp discontinuities in the landscape should be incorporated into the mapping process to allow better interpretation; high-resolution data sources mean that such mapping is increasingly achieved via 'semi-automated' analysis of geomorphometry (Anders *et al.* 2011; Bishop *et al.* 2012). Care with surface topographic representation is critical because, at the catchment scale, the 'process' label infers a dominant concern for process over landform that may not be justified: although there is an interdependent relationship between form and process in geomorphology, for most models of contemporary processes, process rates *and types* will actually be controlled by topography (Lane *et al.* 1998a). As topographic representation provides the boundary conditions under which processes operate (e.g. in defining the locations of surface and sub-surface flow convergence and divergence), it will also influence model output (Moore *et al.* 1994). As indicated above, this is especially likely at the catchment scale where necessary simplifications to process models and terrain representation are inevitable in order to achieve a working model. Zhang *et al.* (2009), for instance, investigated the influence of six different DEM source–resolution combinations on the performance of WEPP, noting that DEM resolution and accuracy influence hillslope length, gradient, channel configuration and channel slope in the model, so influencing the spatial distribution of erosion along hillslopes and in channels and ultimately the gross sediment yield at the watershed outlet.

Land cover (and land use)

Land cover depicts the surface types in a catchment as the basis for estimating related catchment information. Information for surface types can be extracted using their wavelength reflectivity and thermal properties from satellite or aerial imagery. Remotely sensed data can provide detailed, complete and frequent coverage of the Earth's surface, covering wide areas, and have been invaluable for resource management, monitoring, mapping and geomorphology (Walsh *et al.* 1998; see also Chapter 6). However, there are inherent difficulties in delineating continuous thematic layers solely with remotely sensed techniques. For example, soils are stored in terms of discrete areas when in reality, the boundaries between soil and land cover types are not rigid or definable (see Burrough and Frank 1996). General issues of particular importance when defining catchment-scale land cover and/or land use include the following:

- *Spatial resolution:* Finer image resolution distinguishing different land covers types is critical in estimating geomorphology attributes such as sediment sources, but makes for large and unwieldy data sets.

- *Spectral resolution:* Increasing spectral resolution or the number of spectral intervals that a sensor can detect allows spectral differences from surface objects to be better resolved, allowing land cover types to be more meaningfully defined in relation to sediment supply. However, higher resolutions may generate higher noise-to-signal ratios.
- *Radiometric resolution:* The number of bits (e.g. 256, 4096 levels or values) within which the data are recorded – higher resolution allows for a better representation of subtle differences.
- *Geographical coverage:* Catchment studies often involve large areas, and therefore many images, partially compensated by increasingly efficient storage and display mechanisms in modern GIS.
- *Temporal coverage:* Frequency of coverage is very important as attributes such as seasonality may be critical in, for instance, distinguishing hardwood from coniferous tree species, and the time required to obtain cloud-free coverage of large areas will influence the time taken to generate land-cover databases. Issues of image registration become critical when dealing with change detection analyses that rely on data overlays (Sundaresan *et al.* 2007).

Whereas land *cover* describes the type of surface material, land *use* describes the nature of the human activities associated with a particular area of land. Land use cannot be derived fully from remotely sensed techniques and requires ancillary 'training' datasets such as zoning of urban areas to discriminate known categories and increase the accuracy of the classification. Although land cover information is frequently sufficient for use in hydrological or vegetation modelling, catchment models in geomorphology are likely also to require land use information (e.g. for interpretative analyses). Data sources of land-cover and land-use information include a wide variety of map- and image-based products, in addition to any directly sampled data. Relevant land-cover digital data include sources such as the STATSGO and SSURGO soil survey data in the United States (Lytle *et al.* 1996) and related resources such as fire return interval data to help define the effects of fire in a catchment (Swanson 1981), and predictive vegetation and land-cover mapping capabilities designed for large areas with sparse field data information (LEMMA, Ohmann and Gregory 2002).

Channel networks and morphology

Studies that link fluvial processes to geomorphological change require realistic representation of the river channel network and, increasingly, of the river channel morphology as the basis for process-based modelling. Creating a structured river channel network is critical as it enables water and sediment entering the channel reach (from upstream or the adjoining side slope) to be routed downstream. Channels can be extracted from high-resolution DEM surfaces based on non-linear diffusion, contributing area and curvature (Passalacqua *et al.* 2010), but channel networks derived from DEMs frequently make use of a combination of raster and vector data: whereas raster data work well in clearly defined river valleys, in lowland areas a rasterized DEM frequently cannot track the upstream contributing area with accuracy (Rieger 1998; Murray 2007), requiring 'sink-filling' algorithms, supplemental high-resolution imagery or vector data for the channel path (Downs and Priestnall 1999). In SHETRAN, where the channel runs along the edge of a cell, new algorithms were required to allow automatic river network generation (Birkinshaw 2010). A related and critical decision is to determine reliable thresholds based on slope and drainage area to define where the channel starts (Kirkby 1980; Montgomery and Dietrich 1989; Dietrich and Dunne 1993), especially in applications such as SHETRAN where channel width is set as a function of the upstream contributing area (Birkinshaw 2010). Having established the river network, there are frequently significant challenges in characterizing significant features of geomorphological interest over the network extent. Recent developments include a method for spatially disaggregating DEM- or digital orthophotograph-based river networks using GIS routines and then re-aggregating segments of the network based on heterogeneity of fluvial features of interest such as stream power, channel sinuosity or lateral activity (Alber and Piégay 2011). Such analysis lends itself to network-scale examination of multi-scale factors controlling attributes of the channel environment, such as floodplain width (Notebaert and Piégay 2013). Detailed information about channel morphology is increasingly being extracted directly from LiDAR-based point cloud data using GIS software or geostatistical methods (Passalacqua *et al.* 2010).

Model performance

Catchment process models in fluvial geomorphology have been developed from various perspectives with the intention of using process knowledge to replicate and understand geomorphological behaviour. There have subsequently been numerous reviews of catchment process modelling in fluvial geomorphology, with many focused on the role of and potential for RCMs (e.g. Merritt *et al.* 2003; Martin and Church 2004; Brasington and Richards 2007; Coulthard *et al.* 2007; Murray 2007; Van de Wiel *et al.* 2007, 2011; Tucker and Hancock 2010). The issues raised are multifarious (see Chapter 17) but, fundamentally, judgements of model performance are complicated by (i) the inherent complexity of the models, leading to issues with accuracy and uncertainty, and (ii) the extensive field data required for performance evaluation. An additional dimension is whether success in model performance is related to prediction in or learning about the 'real world': it is argued, for instance, that the fundamental role of scientific research is to generate more uncertainties as the basis of research innovation (Odoni and Lane 2011) and, as such, RCMs should be used primarily for exploration about how landforms behave rather than for direct prediction (Coulthard *et al.* 2007).

Accuracy and uncertainty

Stemming from matters related to input data representation, it is evident that the apparent accuracy gains brought about by enhanced catchment-scale analytical capabilities resulting from new technology and increasingly higher resolution data *could* be more than offset by additional sources of data inaccuracy and generalization. This may result in better model 'conceptualization', but less accurate models in terms of their output. As Tucker and Hancock (2010, p. 44) note, '… more complexity and detail may not necessarily produce the desired results'. An interesting twist on this matter exists in relation to the 'effect-of-process' basis of RCMs: here it is argued that the modeller's perceptual view of the landform is arguably more important in model development than the prevailing knowledge of process behaviour, making detailed scrutiny of the model 'of little worth' (Odoni and Lane 2011, p. 168).

Generic issues in model accuracy include the need to achieve consistency in parameter derivation (Kirkby 1996), to tackle 'upscaling' problems associated with process representation (Haff 1996; Kirkby 1996; Brooks and Anderson 1998; Kirkby *et al.* 1998; Tucker and Hancock 2010; see also previous sections) and to resolve the 'inverse problem' of increasing parameter uncertainty away from the present day (Yeh 1986; Brooks and Anderson 1998; Van De Wiel *et al.* 2011). Other fundamental issues relate to the concern for representing the panoply of geomorphological processes via a regular gridded (or other consistent) structure and knowledge gaps in our understanding of geomorphological processes: Tucker and Hancock (2010), for instance, highlight issues of grain size distribution in sediment sources, processes of sediment sorting and comminution in transport, horizontal motion of steep faces (e.g. escarpments, river banks), representation of river channel geometry, the dynamics of debris flow and the role of vegetation dynamics in altering geomorphological processes.

The inherent complexity of catchment process models means that there are multiple sources of uncertainty: Lane (2003) highlights six types, namely closure, structural, solution, process, parameter, initialization and validation (summarized in Odoni and Lane 2011, table 9.2), to which Ewen *et al.* (2006) add run-time errors associated with rainfall and other data that drive catchment processes. Isolating the impact of individual contributions to uncertainty is difficult owing to model complexity and imperfect process understanding (Ewen *et al.* 2006) to the extent that uncertainty investigations to date have often focused on the 'easy' target of parameter uncertainty (Odoni and Lane 2011). Part of the problem, of course, stems from our lack of sufficient spatially distributed input data to drive the models and the paucity of space- and time-specific calibration data, but also the extent to which current models depend on the user expertise required to run them (Merritt *et al.* 2003). The uncertainties associated with explicitly representing processes and small space and time spaces serve, in part, to provide a justification for the development of larger scale parameterizations that are associated with RCMs (Murray 2007).

Validation

As highlighted previously (Downs and Priestnall 2003) and despite numerous validation studies in the intervening decade, there is some consensus that advances in catchment process modelling is limited most fundamentally by the lack of data available for performance testing. Such data limitations extend to the fundamentals of rainfall-runoff modelling and resolution is argued to require major programmes of field data collection (Ewen *et al.* 2006). Various types of data can potentially be used for testing catchment process models and their component parts. They include direct observations of (rapid) landform evolution, monitoring of sediment fluxes, scaled experimental models developed under laboratory situations and natural experiments in constrained field situations (Tucker and Hancock 2010). Validation strategies may not be constrained simply to comparison with field data: acknowledging earlier arguments that numerical models cannot be validated conclusively (Oreskes *et al.* 1994; Haff 1996), Nicholas (2005) suggests two other (or additional) approaches for RCMs, namely an internal validation of the sensitivity of the component parts of the model to their parameterization and grid structure (see also Fawcett *et al.* 1995) and comparison with other modelling approaches. The former technique is a logical approach for catchment process models given their internal complexity and numerous sub-models. The latter could also include comparison between numerical models of the same type: for instance, Hancock *et al.* (2010) provide a 10,000 year simulation test between the SIBERIA and CAESAR RCMs. In the same vein, Coulthard *et al.* (2007) suggest that model testing may potentially be achieved using comparison with flow data, historical records of planform change, sedimentological records and flume data.

Where models are used for prediction, for instance, of sediment yield, an important validation issue is *equifinality*, where the same model output can result from many combinations of internal processes (see Beven 1996). Therefore, the right prediction may be obtained without the model's internal mechanisms accurately representing the catchment processes. This has led to calls for multiple response validation (e.g. Brooks and Anderson 1998) where the model is optimized against multiple outputs (Mroczkowski *et al.* 1997), implying extensive programmes of field data collection: as Thornes *et al.* (1996, p. 137) note, '… models are only as good as their capacity to replicate, to an acceptable level, the magnitude, pattern in space and time and character of real world processes'. In geomorphological applications where processes deriving from the topography are routed back to measure their impact on future topography, this may imply decadal time frame commitments to field monitoring. Further, there are issues related to the potential stochasticity of geomorphological processes: for instance, in early experimental plot tests of SHETRAN in Portugal, one event produced extreme sediment production even though its runoff parameters were similar to other events (Bathurst *et al.* 1996).

Because there is little routine monitoring of geomorphological processes world-wide, the most feasible source of data for validation may be comparison with morphological change itself, using technologies such as LiDAR and GPS to achieve rapid, repeat surveying of catchment terrain and/or the river channel (Higgitt and Warburton 1999), conforming with Nicholas's (2005) testing approach-based 'observations of rapid landform evolution'. However achieved, better validation of catchment process models in geomorphology is going to require field observations because, as concluded by Odoni and Lane (2011, p. 170), '… models without observations, whether informal or formal, primary through fieldwork or secondary through tools like remote sensing and archival records, are highly likely to be very poor models indeed'.

8.8 Prospect

The field of catchment process modelling in geomorphology has blossomed in the last decade. There is an increasing range of software tools for exploratory modelling through GIS using readily available digital data sets (see models collected at http://www.joewheaton.org/river-links/models-and-software) and modules have been developed within statistical software packages such as MATLAB, R and Python that can be used by the research community to develop new analytical routines for modelling rivers and landscapes (e.g. Passalacqua *et al*. 2010). The models reviewed in this chapter are all available for free download and many others are also available (see Merritt *et al*. 2003; Tucker and Hancock 2010). Many models have dedicated web sites providing user forums for users to share information, solve issues and notify advances. The US National Center for Earth-surface Dynamics (NCED) has a dedicated 'Desktop Watersheds Integrated Program' designed to advance the field of landscape evolution prediction (including its integration to terrestrial ecosystems, landscapes and land-use dynamics), based explicitly on the use of high-resolution digital topographic data as the foundation for improving process knowledge (NCED 2011).

Catchment process models in fluvial geomorphology all involve the reduction of a dynamic environmental system into a simplified format that allows researchers to comprehend the complexity of the real world (Goodchild *et al*. 1996), albeit introducing uncertainties at every level of the procedure (Burrough *et al*. 1996). It is not surprising, therefore, that catchment models can be approached from several different starting points that include the conceptual, interpretative, empirical and numerical categories covered in this chapter. Owing to the trade-offs involved in developing each of these model types, the most appropriate model depends on its intended purpose (see Table 8.3). For applications where understanding of yearly-to-decadal fluxes or potential for change is required, the successful application of a process-based reductionist model would perhaps be ideal, providing a pseudo-deterministic ('quasi-mechanical'; Martin and Church 2004) output to facilitate a 'what-if' scenario setting. However, such models are potentially very data and time demanding and may not be practicable under the considerable resource pressures that exist in environmental impact scenarios. In such situations, an alternative approach is to use a geographically specific conceptual understanding of catchment process connectivity, using expert geomorphological knowledge to guide an interpretative modelling approach towards a conclusion or to interpret the outputs of an empirical model. This approach may provide considerable richness (Kirkby 1996) in terms of understanding the catchment historical context for reach-scale fluvial system behaviour and the basis for more targeted, detailed analysis, but may not inherently provide a significant scientific learning experience. Where the geomorphology learning experience is paramount and links to morphological response are critical, an effect-of-process RCM may be the ideal model for generating realistic patterns of fluvial landform evolution, including over extended periods of time. Such patterns may not be appropriate for problem-driven applications where location-specific data are required; here, the comprehensive empiricism of field-based sediment budget may be the most appropriate approach.

Looking to the future, the distinction between model types may diminish. Perhaps as a consequence of the increasing availability of high-resolution catchment data on increasingly powerful PCs with better GIS functionality, physically based gridded modelling will become increasingly accessible. Improved knowledge of geomorphological processes and transfer functions will give the component parts of catchment models increasingly more process accuracy. Interpretative approaches such as watershed analyses already use physically based models such as TOPMODEL and SHALSTAB as component inputs (e.g. Reid *et al*. 2007), and empirical models such as sediment budgets now have a simulation-driven alternative in SedNet (Prosser *et al*. 2001). Reductionist models intended originally for predictive purposes are also being used to explore process understanding; for instance, in SHETRAN, assessments have been undertaken to understand better the relation between rainfall intensity and duration as triggers of shallow landslides (Bovolo and Bathurst 2012) and to investigate potential variability in sediment yield with catchment area free from the need for extensive field monitoring (Birkinshaw and Bathurst 2006). Also, recent developments in the reach-scale capabilities of the CAESAR RCM include the adoption of a two-dimensional hydrodynamic model, Lisflood, that should permit better predictions through increasing process accuracy, perhaps indicative of a trend whereby RCMs become 'more complex' as computing power and better algorithms are developed (Coulthard *et al*. 2007). Such increasingly reductionist-style capacity may typify future *conceptual* advances in RCMs (and other models), along with *structural* advances in dealing with uncertainty, parameter sensitivity and up- and downscaling, and *technological* advances in computing speed, storage capacity and data available through remote sensing (Van De Wiel *et al.* 2011).

Table 8.3 Comparison of selected characteristics of catchment process models by category.

Category and approach	Model requirements			Output utility		
	Catchment process representation	**Data**	**Computing and resources**	**Ability to represent morphological change**	**Prediction**	**Focus of learning**
Interpretative						
Fluvial audit/River Styles®/watershed analysis (WA)						
	Province-specific conceptual model, expert judgement	Field reconnaissance, digital data layers	Low–medium: standard GIS capability; component process models (WA)	Use expert interpretation to integrate historical factors	Requires expert interpretation, fits into decision-making framework	Problem-driven, catchment-specific functioning, process stratification (WA)
Empirical						
Data-driven sediment budgets						
	Province-specific conceptual model, expert judgement	Field monitoring, reconnaissance, GIS for extrapolation	Low: none, or standard GIS capability	Interpret from categorized time periods	Requires expert interpretation	Problem- or science-driven, catchment-specific transfer fluxes
Equation-driven sediment budgets, e.g. *SedNet*						
	Province-specific conceptual model, process equations	Literature or field-based parameterization	Medium	Interpret from categorized time periods	Requires expert interpretation	Problem- or science-driven, catchment-specific transfer fluxes
Numerical						
Process-based distributed modelling, e.g. *WEPP, SHETRAN*						
	Deterministic process equations and transfer rules; upscaling routines	Literature or field-based parameterization	Very high: powerful computers; expert teams for process equations	Short-term erosion and transport processes	Location-specific output; small area scenario simulation capabilities	Problem- or science-driven: flux prediction and/or adequacy of process equations
Effect-of-process based RCMs, e.g. *CHILD, CAESAR*						
	Gridded process simulation and transfer rules, variable for scale	Literature- or field-based parameterization	High: powerful computers; small groups for process transfers	Short- and long-term hillslope and channel evolution	Terrain-typical output; scenario simulation capabilities	Science-driven: process-form interaction, adequacy of process transfer functions

Catchment process modelling in fluvial geomorphology is likely to evolve rapidly over the next decade. Model users will need to be cognizant of different approaches and be prepared to vary their approach to different problems accordingly, but model development in general seems likely to move towards process-based, gridded algorithms that allow the evolution of fluvial morphodynamics to be modelled as a response to geomorphological processes. Various generic issues in geomorphological modelling are not going to disappear (see Chapter 17) (Odoni and Lane 2010, 2011): physically based models will still give 'unphysical' results as a function of scaling issues and large grid sizes (Ewen *et al*. 2000) and the issue of process transfers between gridded cells will perhaps become more acute. Concern for the impact of topography on process will require testing of the effect of altering 30 m sided grids from national DEMs into decimetre grids generated from airborne LiDAR (see Zhang *et al*. 2009). New process understanding coded as algorithms will improve modelling accuracy, although it is probably worth recalling the assertion of Kirkby (1996, p. 263) that 'In geomorphology, few models rise far above empiricism and most 'physically based' models are simply pushing the level of empiricism one level further down.' Logically, this should result in an increasing requirement for field data suitable for validation purposes met as an integral part of funding for model development. The role of catchment process models will also change as it becomes increasingly important to integrate geomorphology processes within ecosystem dynamics as a 'problem-initiated' contribution towards interdisciplinary understanding of ecosystem services.

Acknowledgements

We would like to thank Tim Absalom of the GeoMapping Unit at Plymouth University for drafting Figures 8.1, 8.3 and 8.7 and Elaine Watts of the Cartographic Unit at the University of Nottingham for drafting Figures 8.2, 8.5 and 8.6. Gary Priestnall's contribution to the first edition of this chapter is gratefully acknowledged.

References

Abbott, M.B., Bathurst, J.C., Cunge, J.A, O'Connell, P.E. and Rasmussen, J. 1986a. An introduction to the European Hydrological System – Système Hydrologique Européen, 'SHE', 1: history and philosophy of a physically-based, distributed modelling system. *Journal of Hydrology* 87: 45–59.

Abbott, M.B., Bathurst, J.C., Cunge, J.A, O'Connell, P.E. and Rasmussen, J. 1986b. An introduction to the European Hydrological System – Système Hydrologique Européen, 'SHE', 2: structure of a physically-based, distributed modelling system. *Journal of Hydrology* 87: 61–77.

Ahnert, F. 1976. Brief description of a comprehensive three-dimensional process-response model of landform development. *Zeitschrift für Geomorphologie* 25: 29–49.

Alber, A. and Piégay, H. 2011. Spatial disaggregation and aggregation procedures for characterizing fluvial features at the network-scale: application to the Rhône basin (France). *Geomorphology* 125: 343–360.

Anders, N.S., Seijmonsbergen, A.C. and Bouten, W. 2011. Segmentation optimization and stratified object-based analysis for semi-automated geomorphological mapping. *Remote Sensing of Environment* 115: 2976–2985.

Anderson, M.G., ed. 1988. *Modelling Geomorphological Systems*, Chichester: John Wiley & Sons.

Anderson, J.E., Plourde, L.C. Martin, M.E. Braswell, B.H. Smith, M.-L., Dubayah, R.O., Hofton, M.A. and Blair, J.B. 2008. Integrating waveform lidar with hyperspectral imagery for inventory of a northern temperate forest. *Remote Sensing of Environment* 112: 1856–1870.

Anderson, M.G. and Sambles, K.M. 1988. A review of the bases of geomorphological modelling. In: Anderson, M.G., ed., *Modelling Geomorphological Systems*, Chichester: JohnWiley & Sons, pp. 1–32.

Ascough, J.C. II,, Baffaut, C., Nearing, M.A. and Liu, B.Y. 1997. The WEPP watershed model: I. Hydrology and erosion. *Transactions of the American Society of Agricultural Engineers* 40: 921–933.

Baffaut, C., Nearing, M.A., Ascough, J.C. II, and Liu, B.Y. 1997. The WEPP watershed model: II. Sensitivity analysis and discretization on small watersheds. *Transactions of the American Society of Agricultural Engineers* 40: 935–943.

Baker, V.R. 1996. Hypotheses and geomorphological reasoning. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 57–85.

Batelaan, O., Wang, Z. and De Smedt, F. 1996. An adaptive GIS toolbox for hydrological modelling. In: Kovar, K. and Nachtnebel, H.P., eds., *Application of Geographic Information Systems in Hydrology and Water Resources Management, IAHS Publication 235*, Wallingford: International Association of Hydrological Sciences, pp. 3–10.

Bathurst, J.C., Kilsby, C. and White, S. 1996. Modelling the impacts of climate change and land-use change on basin hydrology and soil erosion in Mediterranean Europe. In: Brandt, C.J. and Thornes, J.B., eds., *Mediterranean Desertification and Land Use*, Chichester: John Wiley & Sons, pp. 355–387.

Bathurst, J.C., Moretti, G., El-Hames, A., Moaven-Hashemi, A. and Burton, A. 2005. Scenario modelling of basin-scale, shallow landslide sediment yield, Valsassina, Italian Southern Alps. *Natural Hazards and Earth System Sciences* 5: 189–202.

Bathurst, J.C., Moretti, G., El-Hames, A., Beguería, S. and García-Ruiz, J.M. 2007. Modelling the impact of forest loss on shallow landslide sediment yield, Ijuez river catchment, Spanish Pyrenees. *Hydrology and Earth System Sciences* 11: 569–583.

Beven, K.J. 1996. Equifinality and uncertainty in geomorphological modelling. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 289–313.

Beven, K.J. and Kirkby, M.J. 1979. A physically based variable contributing-area model of catchment hydrology. *Hydrological Science Bulletin* 24: 43–49.

Birkinshaw, S.J. 2010. Automatic river network generation for a physically-based river catchment model. *Hydrology and Earth System Sciences* 14: 1767–1771.

Birkinshaw, S.J. and Bathurst, J.C. 2006. Model study of the relationship between sediment yield and river basin area. *Earth Surface Processes and Landforms* 31: 750–761.

Birkinshaw, S.J., James, P. and Ewen, J. 2010. Graphical user interface for rapid set-up of SHETRAN physically-based river catchment model. *Environmental Modelling & Software* 25: 609–610.

Bishop, M.P, James, L.A., Shroder, J.F. Jr., and Walsh, S.J. 2012, Geospatial technologies and digital geomorphological mapping: concepts, issues and research. *Geomorphology* 137: 5–26.

Booth, D.B., Leverich, G., Downs, P.W., Dusterhoff, S.R. and Araya, S. 2014. A method for spatially explicit representation of sub-watershed sediment yield, southern California, USA. *Environmental Management* 53: 968–984.

Bovolo, C.I. and Bathurst, J.C. 2012. Modelling catchment-scale shallow landslide occurrence and sediment yield as a function of rainfall return period. *Hydrological Processes* 26: 579–596.

Brasington, J. and Richards, K. 2007. Reduced-complexity, physically-based geomorphological modelling for catchment and river management. *Geomorphology* 90: 171–177.

Brierley G.J. and Fryirs K. 2005. *Geomorphology and River Management: Applications of the River Styles Framework*. Oxford: Blackwell.

Brooks, S.M. and Anderson, M.G. 1998. On the status and opportunities for physical process modelling. In: Longley, P.A., Brooks, S.M., McDonnell, R. and Macmillan, B., eds., *Geocomputation: a Primer*, Chichester: John Wiley & Sons, pp. 193–230.

Burrough, P.A. and Frank, A.U. 1996. *Geographic Objects with Indeterminate Boundaries*. London: Taylor & Francis.

Burrough, P.A. and McDonnell, R.A. 1998. *Principles of Geographical Information Systems*. Oxford: Oxford University Press.

Burrough, P.A., van Rijn, R. and Rikken, M. 1996. Spatial data quality and error analysis issues: GIS functions and environmental modelling. In: Goodchild, M.F., Steyaert, L.T., Parks, B.O., Johnson, C., Maidment, D., Crane, M. and Glendinning, S., eds., *GIS and Environmental Modeling: Progress and Research Issues*, Fort Collins, CO: GIS World Books, pp. 29–34.

Codilean, A.T., Bishop, P. and Hoey, T. 2006. Surface process models and the links between tectonics and topography. *Progress in Physical Geography* 30: 307–333.

Coulthard, T.J., Kirkby, M.J. and Macklin, M.G. 1999. Modelling the impacts of Holocene environmental change in an upland river catchment, using a cellular automaton approach. In: Brown, A.G. and Quine, T.A., eds., *Fluvial Processes and Environmental Change*, Chichester: John Wiley & Sons, pp. 1–46.

Coulthard, T.J., Kirkby, M.J. and Macklin, M.G. 2000. Modelling geomorphic response to environmental change in an upland catchment. *Hydrological Processes* 14: 2031–2045.

Coulthard, T.J., Macklin, M.G. and Kirkby, M.J. 2002. A cellular model of Holocene upland river basin and alluvial fan evolution. *Earth Surface Processes and Landforms* 27: 269–288.

Coulthard, T.J., Hicks, M.D. and Van De Wiel, M.J. 2007. Cellular modelling of river catchments and reaches: advantages, limitations and prospects. *Geomorphology* 90: 192–207.

Davis, W.M. 1899. The geographical cycle. *Geographical Journal* 14: 481–504.

Dietrich, W.E. and Dunne, T. 1978. Sediment budget for a small catchment in mountainous terrain. *Zeitschrift für Geomorphologie*, 29(Suppl.): 191–206.

Dietrich, W.E. and Dunne, T. 1993. The channel head. In: Beven, K.J. and Kirkby, M.J., eds., *Channel Network Hydrology*, Chichester: John Wiley & Sons, pp. 175–219.

Dietrich, W.E., Dunne, T., Humphrey, N.F. and Reid, L.M. 1982. Construction of sediment budgets for drainage basins. In: Swanson, F.J., Janda, R.J., Dunne, T. and Swanston, D.N., eds., *Workshop on Sediment Budgets and Routing in Forested Drainage Basins, General Technical Report PNW-141*,USDA Forest Service Pacific Northwest Forest and Range Experiment Station, Portland, OR: pp. 5–23.

Dietrich, W.E., Bellugi, D.G., Sklar, L.S., Stock, J.D., Heimsath, A.M. and Roering J.J. 2003. Geomorphic transport laws for predicting landscape form and dynamics. In: Wilcock, P.R. and Iverson, R.M., eds., *Prediction in Geomorphology, Geophysical Monograph 135*, Washington, DC: American Geophysical Union, pp. 103–132.

Downs, P.W. and Booth, D.B. 2011. Geomorphology in environmental management. In: Gregory, K.J. and Goudie, A.S., eds., *The SAGE Handbook of Geomorphology*, London: SAGE Publications, pp. 78–104.

Downs, P.W. and Gregory, K.J. 2004. *River Channel Management: Towards Sustainable Catchment Hydrosystems*, London: Arnold.

Downs, P.W. and Priestnall, G. 1999. System design for catchment-scale approaches to studying river channel adjustments using a GIS. *International Journal of Geographical Information Systems* 13: 247–266.

Downs, P.W. and Priestnall, G. 2003. Modelling catchment processes. In: Kondolf, G.M. and Piégay, H., eds., *Tools in Fluvial Geomorphology*, 1st edn., Chichester: John Wiley & Sons, pp. 205–230.

Downs, P.W. and Thorne, C.R. 1996. The utility and justification of river reconnaissance surveys. *Transactions of the Institute of British Geographers* 21: 455–468.

Dun, S., Wu, J.Q., Elliot, W.J., Robichaud, P.R., Flanagan, D.C., Frankenberger J.R., Brown, R.E. and Xu, A.C. 2009. Adapting the Water Erosion Prediction Project (WEPP) model for forest applications. *Journal of Hydrology* 466: 46–54.

Dunn, S.M., Mackay, R., Adams, R. and Oglethorpe, D.R. 1996. The hydrological component of the NELUP decision-support system: an appraisal. *Journal of Hydrology* 177: 213–235.

Ewen, J., Parkin, G. and O'Connell, P.E. 2000. SHETRAN: distributed river basin flow and transport modeling system. *Proceedings of the American Society of Civil Engineers, Journal of Hydrologic Engineering* 5: 250–258.

Ewen, J., O'Donnell, G., Burton, A. and O'Connell, E. 2006. Errors and uncertainty in physically-based rainfall-runoff modelling of catchment change effects. *Journal of Hydrology* 330: 641–650.

Fawcett, K.R., Anderson, M.G., Bates, P.D., Jordan, J.-P. and Bathurst, J.C. 1995. The importance of internal validation in the assessment of physically-based distributed models. *Transactions of the Institute of British Geographers* 20: 248–265.

Flanagan, D.C., Nearing, M.A. and Laflen, J.M., eds. 1995. *USDA – Water Erosion Prediction Project: Hillslope Profile and Watershed Model Documentation, US Department of Agriculture Agricultural Research Service National Soil Erosion Research Laboratory Report 10*, West Lafayette, IN: US Department of Agriculture.

Flanagan, D.C., Gilley, J.E. and Franti, T.G. 2007. Water Erosion Prediction Project (WEPP): development history, model capabilities and future enhancements. *Transactions of the American Society of Agricultural and Biological Engineers* 50: 1603–1612.

Fonstad M.A. 2006. Cellular automata as analysis and synthesis engines at the geomorphology–ecology interface. *Geomorphology* 77: 217–234.

Foster, G.R. and Lane, L.J. 1987. *User Requirements: USDA – Water Erosion Prediction Project WEPP. United States Department of Agriculture Agricultural Research Service National Soil Erosion Research Laboratory Report 1*, West Lafayette, IN: US Department of Agriculture.

Glennie, C., Brooks, B., Ericksen, T., Hauser, D., Hudnut, K., Foster J. and Avery J. 2013. Compact multipurpose mobile laser scanning system – initial tests and results. *Remote Sensing* 5: 521–538.

Goodchild, M.F., Steyaert, L.T., Parks, B.O., Johnson, C., Maidment, D., Crane, M. and Glendinning, S., eds. 1996. *GIS and Environmental Modeling: Progress and Research Issues*, Fort Collins, CO: GIS World Books.

Gregory, K.J. 2000. *The Changing Nature of Physical Geography*, London: Arnold.

Gregory, K.J., Benito, G. and Downs, P.W. 2008. Applying fluvial geomorphology to river channel management: background for progress towards a palaeohydrology protocol. *Geomorphology* 98: 153–172.

Haff, P.K. 1996. Limitations on predictive modelling in geomorphology. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 337–360.

Hancock, G.R., Lowry, J.B.C., Coulthard, T.J., Evans, K.G. and Moliere, D.R. 2010. A catchment scale evaluation of the SIBERIA and CAESAR landscape evolution models. *Earth Surface Processes and Landforms* 35: 863–875.

Heritage, G. and Hetherington, D. 2007. Towards a protocol for laser scanning in fluvial geomorphology. *Earth Surface Processes and Landforms* 32: 66–74.

Higgitt, D.L. and Warburton, J. 1999. Applications of differential GPS in upland fluvial geomorphology. *Geomorphology* 29: 121–134.

Jaafar, J., Priestnall, G., Mather, P.M. and Vieira, C.A. 1999. Construction of DEM from LiDAR DSM using morphological filtering, conventional statistical approaches and artificial neural networks. In: *Earth Observation: from Data to Information. Proceedings of the 25th International Conference of the Remote Sensing Society (RSS'99), University of Wales at Cardiff and Swansea*, Nottingham: Remote Sensing Society, pp. 299–306.

Kellerhals, R., Church, M. and Bray, D.I. 1976. Classification and analysis of river processes. *Journal of the Hydraulics Division, American Society of Civil Engineers* 102: 813–829.

King, L. 1962. *The Morphology of the Earth*, Edinburgh: Oliver & Boyd.

Kirkby, M.J. 1971. Hillslope process-response models based on the continuity equation. *Special Publication, Institute of British Geographers* 3: 15–30.

Kirkby, M.J. 1980. The stream head as a significant geomorphic threshold. In: Coates, D.R and Vitel, A.D., eds., *Thresholds in Geomorphology*, London: Allen and Unwin, pp. 53–73.

Kirkby, M.J. 1996. A role for theoretical models in geomorphology? In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 257–272.

Kirkby, M.J., Abrahart, R., McMahon, M.D., Shao, J. and Thornes, J.B. 1998. MEDALUS soil erosion models for global change. *Geomorphology* 24: 35–49.

Knighton, A.D. 1998. *Fluvial Forms and Processes: a New Perspective*, London: Arnold.

Kondolf, G.M. and Matthews, W.V.G. 1991. Unmeasured residuals and sediment budgets: a cautionary note. *Water Resources Research* 27: 2483–2486.

Lane, L.J. and Nearing, M.A., eds. 1989. *USDA – Water Erosion Prediction Project: Hillslope Profile Model Documentation, United States Department of Agriculture Agricultural Research Service National Soil Erosion Research Laboratory Report 2*, West Lafayette, IN: US Department of Agriculture.

Lane, S.N. 1998. Hydraulic modelling in hydrology and geomorphology: a review of high resolution approaches. *Hydrological Processes* 12: 1131–1150.

Lane, S.N. 2003. Numerical modelling in physical geography: understanding, explanation and prediction. In: Clifford, N.J. and Valentine, G., eds., *Key Methods in Geography*, London: SAGE Publications, pp. 263–290.

Lane, S.N., Richards, K.S. and Chandler, J.H. 1998a. Landform monitoring, modelling and analysis: landform in geomorphological research. In: Lane, S.N., Richards, K.S. and Chandler, J.H., eds., *Landform Monitoring, Modelling and Analysis*, Chichester: John Wiley & Sons, pp. 1–18.

Lane, S.N., Richards, K.S. and Chandler, J.H., eds. 1998b. *Landform Monitoring, Modelling and Analysis*, Chichester: John Wiley & Sons.

Lytle, D.J., Bliss, N.B. and Waltman, S.W. 1996. Interpreting the State Soil Geographic Database (STATSCO). In: Goodchild, M.F., Steyaert, L.T., Parks, B.O., Johnson, C., Maidment, D., Crane, M. and Glendinning, S., eds., *GIS and Environmental Modeling: Progress and Research Issues*, Fort Collins, CO: GIS World Books, pp. 49–52.

Martin, Y. and Church, M. 2004. Numerical modelling of landscape evolution: geomorphological perspectives. *Progress in Physical Geography* 28: 317–339.

McCullagh, M.J. 1998. Quality, use and visualisation in terrain modelling. In: Lane, S.N., Richards, K.S. and Chandler, J.H., eds., *Landform Monitoring*, Modelling and Analysis, Chichester: John Wiley & Sons, pp. 95–118.

McKean, J., Nagel, D., Tonina, D., Bailey, P., Wright, W.C., Bohn, C. and Nayegandhi, A. 2009. Remote sensing of channels and riparian zones with a narrow-beam aquatic–terrestrial LIDAR. *Remote Sensing* 1: 1065–1096.

Merritt, W.S., Letcher, R.A. and Jakeman, A.J. 2003. A review of erosion and sediment transport models. *Environmental Modelling & Software* 18: 761–799.

Montgomery, D.R. 1999. Process domains and the river continuum. *Journal of the American Water Resources Association* 35: 397–410

Montgomery, D.R. and Dietrich, W.E. 1989. Source areas, drainage density and channel initiation. *Water Resources Research* 25: 1907–1918.

Montgomery, D.R. and Dietrich, W.E. 1994. A physically based model for the topographic control on shallow landsliding. *Water Resources Research* 30: 1153–1171.

Montgomery, D.R. and Foufoula-Georgiou, E. 1993. Channel network source representation used digital elevation models. *Water Resources Research* 29: 3925–3934.

Montgomery, D.R., Grant, G.E. and Sullivan, K. 1995. Watershed analysis as a framework for implementing ecosystem management. *Water Resources Bulletin* 31: 369–386.

Montgomery, D.R., Dietrich, W.E. and Sullivan, K. 1998. The role of GIS in watershed analysis. In: Lane, S.N., Richards, K.S. and Chandler, J.H., eds., *Landform Monitoring, Modelling and Analysis*, Chichester: John Wiley & Sons, pp. 241–261.

Moore, I.D., Grayson, R.B. and Ladson, A.R. 1994. Digital terrain modelling: a review of hydrological, geomorphological and biological applications. In: Beven, K.J. and Moore, I.D., eds., *Terrain Analysis and Distributed Modelling in Hydrology: Advances in Hydrological Processes*, Chichester: John Wiley & Sons, pp. 7–34.

Moorhead, R., Hood, R. and Coffey, J. 2012. *Optimal Unmanned Aircraft Systems River Observing Strategy Workshop, February 21–23, 2012, Boulder, Colorado. Summary Report.* http://uas.noaa.gov/library/workshop/UAS_RFC_Workshop_Summary_final.pdf (accessed 8 October 2015).

Mroczkowski, M., Raper, G.P. and Kuczera, G. 1997. The quest for a more powerful validation of conceptual catchment models. *Water Resources Research* 33: 2325–2336.

Murray, A.B. 2007. Reducing model complexity for explanation and prediction. *Geomorphology* 90: 178–191.

Murray, A.B. and Paola, C. 1994. A cellular model of braided rivers. *Nature* 371: 54–57.

NCED (National Center for Earth-surface Dynamics). 2011. *Annual Report 2011. IIa. Desktop Watersheds (Watersheds) Integrated Program*, pp. 15–38. http://www.nced.umn.edu/system/files/NCED_AR_2011-Watershedsportion.pdf (accessed 8 October 2015).

Nicholas, A.P. 2005. Cellular modelling in fluvial geomorphology. *Earth Surface Processes and Landforms* 30: 645–649.

Nicholas, A.P. and Quine, T.A. 2007. Crossing the divide: representation of channels and processes in reduced-complexity river models at reach and landscape scales. *Geomorphology* 90: 318–339.

Notebaert, B. and Piégay, H. 2013. Multi-scale factors controlling the pattern of floodplain width at a network scale: the case of the Rhône basin, France. *Geomorphology* 200: 155–171.

Odoni, N.A. and Lane, S.N. 2010. Knowledge-theoretic models in hydrology. *Progress in Physical Geography* 34: 151–171.

Odoni, N.A. and Lane, S.N. 2011. The significance of models in geomorphology: from concepts to experiments. In: Gregory, K.J. and Goudie, A.S., eds., *The SAGE Handbook of Geomorphology*, London: SAGE Publications, pp. 154–173.

Ohmann, J.L. and Gregory, M.J. 2002. Predictive mapping of forest composition and structure with direct gradient analysis and nearest neighbor imputation in coastal Oregon, U.S.A. *Canadian Journal of Forest Research* 32: 725–741.

Oreskes, N., Shrader-Frechette, K. and Belitz, K. 1994. Verification, validation and confirmation of numerical models in the earth sciences. *Science* 264: 641–646.

Paola, C. 2001. Modelling stream braiding over a range of scales. In: Mosley, M.P., ed., *Gravel Bed Rivers V*, Wellington: New Zealand Hydrological Society, pp. 11–46.

Passalacqua, P., Do Trung, T., Foufoula-Georgiou, E., Sapiro, G. and Dietrich W.E. 2010. A geometric framework for channel network extraction from LiDAR: nonlinear diffusion and geodesic paths. *Journal of Geophysical Research*, 115: F01002.

Penck, W. 1953. *Morphological Analysis of Landforms, transl*. Czech, H. and Boswell, K.C., London: Macmillan.

Perron, J.T. and Fagherazzi, S. 2012. The legacy of initial conditions in landscape evolution. *Earth Surface Processes and Landforms* 37: 52–63.

Prosser, I.P., Rustomji, P., Young, B., Moran, C. and Hughes, A. 2001. *Constructing river basin sediment budgets for the National Land and Water Resources Audit, Technical Report 15/01*, Canberra: CSIRO Land and Water, 34pp.

Reid, L.M. 1993. *Research and Cumulative Watershed Effects, General Technical Report PSW-GTR-141*, Albany, CA: USDA Forest Service, Pacific Southwest Research Station, 118pp.

Reid, L.M. and Dunne, T. 1996. *Rapid Evaluation of Sediment Budgets*, Reiskirchen: Catena.

Reid, L.M. and McCammon, B.P. 1993. A Procedure for Watershed Analysis. Interagency Forest Ecosystem Management Assessment Team, USDA Forest Service Region 6.

Reid, S.C., Lane, S.N., Montgomery, D.R. and Brookes, C.J. 2007. Does hydrological connectivity improve modelling of coarse sediment delivery in upland environments? *Geomorphology* 90: 263–282.

Reitberger, J., Krzystek, P. and Stilla, U. 2009. Benefit of airborne full waveform LIDAR for 3D segmentation and classification of single trees. Presented at the ASPRS Annual Conference, Baltimore, March 9–13, 2009. http://www.asprs.org/a/publications/proceedings/baltimore09/0078.pdf (accessed 8 October 2015).

Rieger, W. 1998. A phenomenon-based approach to upslope contributing area and depressions in DEMs. *Hydrological Processes* 12: 857–872.

Rosette, J.C., Field, R.N., Nelson, R.F., Decola, P., Cook, B.D. and Degnan, J.J. 2011. Single-photon LiDAR for vegetation analysis. Presented at the American Geophysical Union Fall Meeting 2011, abstract B51C-0415. http://adsabs.harvard.edu/abs/2011AGUFM.B51C0415R (accessed 8 October 2015).

Schumm, S. A. 1991. *To Interpret the Earth: Ten Ways to Be Wrong*. Cambridge: Cambridge University Press.

Schumm, S.A. and Lichty, R.W. 1965. Time, space and causality in geomorphology. *American Journal of Science*, 263: 110–119.

Sear, D.A., Newson, M.D. and Brookes, A. 1995. Sediment-related river maintenance: the role of fluvial geomorphology. *Earth Surface Processes and Landforms* 20: 629–647.

Sear, D.A., Newson, M.D., Thorne, C.R. 2003. *Guidebook of Applied Fluvial Geomorphology, R and D Technical Report FD 1914*, London: Defra/Environment Agency.

Sear, D.A., Newson, M.D., Hill, C., Old, J. and Branson, J. 2009. A method for applying fluvial geomorphology in support of catchment-scale river restoration planning. *Aquatic Conservation: Marine and Freshwater Ecosystems* 19: 506–519.

Slaymaker, O. 2003. The sediment budget as conceptual framework and management tool. *Hydrobiologica* 494: 71–82.

Slaymaker, O. 2008. The future of geomorphology. *Geography Compass* 3: 329–349.

Smith, J.S. and Smith, D.G. 1996. Operational experiences of digital photogrammetric systems. *International Archives of Photogrammetry and Remote Sensing* 31(Part B2): 357–362.

Smith, M.J., Smith, D.G., Tragheim, G. and Holt, M. 1997. DEMs and ortho-images from aerial photographs. *Photogrammetric Record* 15: 945–950.

Sundaresan, A., Varshney, P.K. and Arora, M.K. 2007. Robustness of change detection algorithms in the presence of registration errors. *Photogrammetric Engineering and Remote Sensing* 73: 375–383.

Swanson, F.J. 1981. Fire and geomorphic processes. In: Mooney, H., Bonnicksen, T.H., Christensen, N.L., Lotan, J.E. and Reiners, W.A., eds., *Proceedings of the Conference on Fire Regimes and Ecosystem Properties, December 11–15, 1978, Honolulu, Hawaii, General Technical Report WO-26*, Washington, DC: USDA Forest Service, pp. 410–420.

Thorne, C.R. 1997. Channel types and morphological classification. In: Thorne, C.R., Hey, R.D. and Newson, M.D. eds., *Applied Fluvial Geomorphology in River Engineering Management*, Chichester: John Wiley & Sons, pp. 175–222.

Thorne, C.R. 1998. *Stream Reconnaissance Handbook*, Chichester: John Wiley & Sons.

Thornes, J.B., Shao, J.X., Diaz, E., Roldan, A., McMahon, M. and Hawkes, J.C. 1996. Testing the MEDALUS hillslope model. *Catena* 26: 137–160.

Trimble, S.W. 1999. Decreased rates of alluvial sediment storage in the Coon Basin, Wisconsin, 1975–93. *Science* 285: 1244–1246.

Trimble, S.W. 2009. Fluvial processes, morphology and sediment budgets in the Coon Creek Basin, WI, USA, 1975–1993. *Geomorphology* 108: 8–23.

Tucker, G.E. and Hancock, G.R. 2010. Modelling landscape evolution. *Earth Surface Processes and Landforms* 35: 28–50.

Tyler, D.J. and Greenlee, S.K. 2012. *Creation of Digital Contours That Approach the Characteristics of Cartographic Contours, Scientific Investigations Report 2012-5167*, Reston, VA: US Geological Survey.

Van De Wiel, M.J., Coulthard, T.J., Macklin, M.G. and Lewin, J. 2007. Embedding reach-scale fluvial dynamics within the CAESAR cellular automaton landscape evolution model. *Geomorphology* 90: 283–301.

Van De Wiel, M.J., Coulthard, T.J., Macklin, M.G. and Lewin, J. 2011. Modelling the response of river systems to environmental change: progress, problems and prospects for palaeo-environmental reconstructions. *Earth Science Reviews* 104: 167–185.

Wallace, L., Lucieer, A., Watson, C. and Turner, D. 2012. Development of a UAV-LiDAR system with application to forest inventory. *Remote Sensing* 4: 1519–1543.

Walsh, S.J., Butler, D.R. and Malanson, G.P. 1998. An overview of scale, pattern, process relationships in geomorphology: a remote sensing and GIS perspective. *Geomorphology* 21: 183–205.

Wang, L., Wu, J.Q. Elliot, W.J., Dun, S., Lapin, S. Fiedler, F.R. and Flanagan, D.C. 2009. Implementation of channel-routing routines in the Water Erosion Prediction Project (WEPP) model. In: *Proceedings of the Fourth SIAM Conference on Mathematics for Industry, 9–10 October 2009*, San Francisco, Society for Industrial and Applied Mathematics. Available at: http://www.siam.org/proceedings/industry/2009/mi09_014_wangl.pdf (accessed 8 October 2015).

Warburton, J. 2011. Sediment transport and deposition. In: Gregory, K.J. and Goudie, A.S., eds., *The SAGE Handbook of Geomorphology*, London: SAGE Publications, pp. 326–342.

Wasson, R.J. 2002. Sediment budgets, dynamics and variability: new approaches and techniques. In: Dyer, F.J., Thoms, M.C. and Olley, J.M. eds., *The Structure, Function and Management Implications of Fluvial Sedimentary Systems, IAHS Publication 276*, Wallingford: International Association for Hydrological Sciences, pp. 471–478.

Wicks, J.M. and Bathurst, J.C. 1996. SHESED: a physically-based, distributed erosion and sediment yield component for the SHE hydrological modelling system. *Journal of Hydrology* 175: 213–238.

Wilkinson, S., Henderson, A., Chen, Y. and Sherman, B. 2004. *SedNet User Guide, Version 2. Client Report*, Canberra: CSIRO Land and Water. www.toolkit.net.au/sednet (accessed 8 October 2015).

Wilkinson, S.N., Prosser, I.P., Rustomji, P. and Read, A.M. 2009. Modelling and testing spatially distributed sediment budgets to relate erosion processes to sediment yields. *Environmental Modelling and Software* 24: 489–501.

Willgoose, G., Bras, R.L., Rodríguez-Iturbe, I. 1991. A coupled channel network growth and hillslope evolution model. 1. Theory. *Water Resources Research* 27: 1671–1684.

Williams, J.D., Gollany, H.T., Siemens, M.C., Wuest, S.B. and Long, D.S. 2009. No-till and inversion tillage practices effect on crop yields and soil and water conservation in the rainfed croplands North-Central Oregon. *Journal of Soil and Water Conservation* 64: 43–52.

Williams, J.D., Dun, S., Robertson, D.S., Wu, J.Q., Brooks, E.S., Flanagan, D.C. and McCool, D.K. 2010. WEPP simulations of dryland cropping systems in small drainages of northeastern Oregon. *Journal of Soil and Water Conservation* 65: 22–33.

Wilson, J.P., Lam, C.S. and Deng, Y. 2007. Comparison of the performance of flow-routing algorithms used in GIS-based hydrologic analysis, *Hydrological Processes* 21: 1026–1044.

Wischmeier, W.H and Smith, D.D. 1978. *Predicting Rainfall Erosion Losses – A Guide to Conservation Planning, Agricultural Handbook 537*, Washington DC: US Department of Agriculture.

Wise, S.M. 1998. The effect of GIS interpolation errors on the use of digital elevation models in geomorphology. In: Lane, S.J., Richards, K.S. and Chandler, J.H., eds., *Landform Monitoring, Modelling and Analysis*, Chichester: John Wiley and Sons, pp. 139–164.

Yeh, W.W-G. 1986. Review of parameter identification procedures in groundwater hydrology: the inverse problem. *Water Resource Research* 22: 95–108.

Zhang, J.X., Wu, J.Q., Chang, K., Elliot, W.J. and Dun S. 2009. Effects of DEM source and resolution on WEPP hydrologic and erosion simulation: a case study of two forest watersheds in northern Idaho. *Transactions of the American Society of Agricultural and Biological Engineers*, 52: 447–457.

SECTION IV

Chemical, Physical and Biological Evidence: Dating, Emphasizing Spatial Structure and Fluvial Processes

CHAPTER 9

Using environmental radionuclides, mineral magnetism and sediment geochemistry for tracing and dating fine fluvial sediments

Des Walling[1] and Ian Foster[2]
[1]University of Exeter, Exeter, UK
[2]University of Northampton, Northampton, UK

9.1 Introduction

The last 3–4 decades have witnessed a significant increase in the number of research papers that have reported the use of measurements of the radionuclide activities, magnetic properties and geochemical signatures of soils and sediments to support geomorphological investigations of the mobilization, transfer and storage of fine sediment in catchments by fluvial processes. Such measurements have been primarily used in three ways: (i) to provide a chronology for deposited sediments, (ii) to estimate rates of soil and sediment redistribution on hillslopes and within catchments over a range of timescales and (iii) to identify the provenance of the fine sediment transported by rivers by comparing the signatures of transported and deposited sediment with those of potential sources.

The tools referred to above have been employed over many different time-scales, ranging from studies of contemporary sediment budgets to reconstruction of long-term landscape evolution. Different cosmogenic radionuclides have, for example, provided a means of estimating long-term rates of landscape lowering through beryllium-10 (^{10}Be) measurements and also short-term rates of soil loss associated with individual storm events through beryllium-7 (^{7}Be) measurements. Similarly, the geochemical properties of fluvial sediment have been used as fingerprints, to establish the contemporary fine sediment sources in a catchment and to reconstruct long-term changes in the provenance of alluvial sediments. Geomorphologists now have access to a wide range of dating techniques (e.g. Walker 2005), including well-established methods such as radiocarbon dating (^{14}C). Although the ^{14}C technique itself is not new, significant improvements in calibrating radiocarbon dates have been made in recent years and the time-scale over which these calibrations have been made has been extended back to around 50,000 radiocarbon years BP (e.g. Fairbanks *et al.* 2005; Reimer *et al.* 2009). New methods, or the refinement of existing methods, for dating quartz and feldspar (e.g. luminescence dating; Wintle and Murray 2006) and for using cosmogenic radionuclides to estimate the age of regolith or exposed rock surfaces have provided geomorphologists with a set of tools that allow the age of landscape surfaces to be determined over different time-scales. This chapter, however, will focus on the use of gamma-emitting radionuclide, mineral magnetic and geochemical measurements to investigate the mobilization, transfer and storage of soil and sediment in catchments, and thus their sediment budgets, over contemporary and short- to medium-term time-scales (i.e. the last ~100 years). These are the main time-scales of interest to river managers when diagnosing sediment problems and planning remediation measures.

The chapter is divided into three sections. The first focuses on the measurements (the tools), the second provides a review of major applications and the third presents a case study where some of the tools have been successfully applied.

9.2 The tools

Gamma-emitting radionuclides

Environmental radioactivity derives from three major sources:

1. Primordial radionuclides, which were created at the time of the formation of the earth (e.g. ^{40}K, ^{235}U, ^{238}U) and which often decay to produce unstable (radioactive) daughter products (e.g. ^{210}Pb).
2. Cosmogenic radionuclides, produced in the upper atmosphere by cosmic ray bombardment [e.g. ^{7}Be, which is formed by the spallation of O and N atoms in the troposphere and stratosphere (Zapata *et al.* 2002; IAEA 2014)].
3. Anthropogenic radionuclides, created as a by-product of nuclear fission from either weapons testing or nuclear accidents such as Chernobyl (e.g. ^{137}Cs).

Table 9.1 lists the most commonly-analysed gamma-emitting radionuclides in soils and sediments. Three of the radionuclides shown in Table 9.1 enter the catchment as atmospheric fallout [^{7}Be, ^{137}Cs and ^{210}Pb$_{un}$ (unsupported ^{210}Pb)]. These

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

Table 9.1 Gamma-emitting radionuclides in environmental samples commonly used in the analysis of soils and sediments.

Isotope and (half-life)	Energy (keV)	Main origin	Secondary origin	Notes
^{210}Pb (22.3 years)	46.5	Atmospheric fallout (unsupported)	^{226}Ra decay (supported)	Atmospheric from ^{222}Rn (radon gas) ^{226}Ra from ^{238}U decay series
^{234}Th† (24.1 days)	63.3	Natural		^{238}U decay series
^{235}U (7.04×10^{8} years)	143 and 186	Natural		^{235}U decay series
^{214}Pb‡ (26.8 min)	295 and 351	Natural		^{238}U/^{226}Ra decay
^{7}Be (53 days)	477	Cosmogenic		Produced continuously in upper atmosphere
^{137}Cs (30 years)	662	Fission: weapons* fallout	Nuclear accidents**	First occurrence 1954*, peaks in 1963* and 1986**
^{228}Ac§ (6.14 h)	338 and 911	Natural		^{232}Th decay series
^{212}Pb¶ (10.6 h)	239	Natural		^{232}Th decay
^{40}K (1.28×10^{9} years)	1461	Natural		Primordial

*Produced global fallout but first occurrence in southern hemisphere in 1956 and peak in 1965 (see text for explanation).
**The Chernobyl nuclear accident produced regional fallout mostly limited to the northern hemisphere.
† ^{234}Th is the immediate daughter of ^{238}U (half-life 4.68×10^{9} years).
‡ ^{214}Pb derives from the decay of ^{226}Ra (half-life 1600 years) through ^{222}Rn (half-life 3.82 days) and ^{218}Po (half-life 3.05 min) and is assumed to be in equilibrium with ^{226}Ra when stored in sealed sample holders for a minimum of 21 days.
§ ^{228}Ac derives from the decay of ^{232}Th (half-life 14.05×10^{9} years) through ^{228}Ra (half-life 5.75 years).
¶ ^{212}Pb derives from further decay of ^{228}Ac. Part of this decay chain results in release of a radioactive gas, ^{220}Ra (thoron; half-life 55.6 s).
Source: Foster and Keay-Bright, 2007. Reproduced with permission of Elsevier.

have a strong affinity for soil and sediment particles in most environments and have proved particularly useful for tracing sediment movement and dating sediment deposits (Matisoff *et al.* 2002; Appleby 2008, 2013; Mabit *et al.* 2008; Blake *et al.* 2009; Walling 2012). ^{7}Be has a very short half-life (the period over which 50% of a nuclide's activity is lost by nuclear decay) of ~53 days. It is produced at a relatively constant rate in the upper atmosphere, giving a relatively constant supply to the catchment surface, when averaged over time. The limited information available, shows annual fallout fluxes ranging between 412 and 6350 Bq m^{-2} (Walling 2012). Fallout is, however, primarily associated with precipitation events and may therefore be characterized by considerable short-term and seasonal variability. This variability, when combined with the short half-life of ^{7}Be, leads to fluctuations in the amount of activity, or the inventory, stored in the surface soil. ^{7}Be is usually found only in the upper few millimetres of an undisturbed soil/sediment surface exposed to fallout, since its short half-life limits the time available for deeper penetration.

In contrast to ^{7}Be, where the supply to the catchment surface is relatively constant from year to year, the fallout of ^{137}Cs was closely linked to the timing and location of 'bomb tests' in the 1950s and early 1960s. Most of these tests took place in the northern hemisphere and the total bomb fallout receipt in the southern hemisphere was only about 10% of that in the northern hemisphere. In the northern hemisphere, fallout commenced around 1954 and reached a peak in 1963, the year of the Nuclear Test Ban treaty. In the absence of subsequent bomb tests, fallout declined rapidly to become negligible by the end of the 1970s (Walling 2012). In the southern hemisphere, fallout was delayed, commencing in 1956 and peaking in 1965 (Rowntree and Foster 2012). In many areas of Europe and adjacent regions, the Chernobyl disaster in 1986 provided a further short-lived ^{137}Cs fallout input. In areas close to Chernobyl this greatly exceeded the earlier bomb fallout. Because of its longer half-life, ^{137}Cs is found at greater depths in undisturbed soil and sediment than ^{7}Be. As a result of slow downward diffusion and bioturbation, ^{137}Cs is commonly found to depths of about 15–20 cm in undisturbed soils (Walling 2012). The depth can increase in cultivated soils, due to mixing, and in depositional sites, where progressive burial of the original surface may occur.

^{210}Pb is a naturally occurring isotope of the ^{238}U decay series, with a half-life of 22.3 years. It derives from the decay of ^{222}Rn (a gas), a daughter of ^{226}Ra, The ^{222}Rn diffuses through the soil profile into the atmosphere, where it decays to produce ^{210}Pb and eventually returns to the ground surface as fallout. The ^{210}Pb falling back to the surface is not in equilibrium with ^{226}Ra and produces an excess with its parent (Robbins 1978). This is referred to as the unsupported or excess ^{210}Pb ($^{210}Pb_{un}$). $^{210}Pb_{un}$ is calculated by measuring the parent isotope (^{226}Ra) to determine the supported ^{210}Pb and subtracting this from the total amount of ^{210}Pb in a sample. ^{226}Ra cannot be measured directly by gamma spectrometry and ^{214}Pb is measured to obtain the ^{226}Ra activity (Murray *et al.* 1987; Gilmore 2008). In contrast to ^{137}Cs, fallout of $^{210}Pb_{un}$ has been relatively constant over time (Crickmore *et al.* 1990). The limited data on annual fallout show substantial spatial variability with rates of between 31 and 840 Bq m^{-2} per year (Walling 2012).

The base unit for reporting radionuclide activities in environmental samples is the becquerel (Bq). This defines the activity of a radionuclide in a sample and is the number of nuclear disintegrations per second (Wallbrink *et al.* 2002). Since the number of disintegrations is also a function of sample mass, activities are conventionally reported in becquerels per kilogram (Bq kg^{-1}). However, geomorphological investigations frequently require a measure of the total activity or inventory per unit area (Bq m^{-2}) contained in a soil or sediment deposit at a particular point. For such measurements, it is important that the core or depth incremental samples should extend over the entire ^{7}Be, $^{210}Pb_{un}$ or ^{137}Cs depth distribution [see Loughran *et al.* (2002) for sampling methods]. The inventory is calculated from measurements of radionuclide activity, sample volume and dry bulk density, although the procedure can be simplified to require only mass and activity, if the core is collected using a core tube of known internal cross-sectional area.

In soil and sediment redistribution investigations, there is frequently a need to establish the local reference inventory, which is the inventory expected in the absence of erosion or deposition. This will directly reflect the fallout input. In the case of ^{137}Cs and $^{210}Pb_{un}$, this value is obtained by sampling a flat, undisturbed area where there is no evidence of erosion or deposition and where there has been no history of cultivation for 50 years or more. Multiple samples are commonly collected from the reference site in order to obtain a reliable estimate of the reference inventory and to characterize the local variability in its magnitude.

Several methods can be used to determine the gamma-emitted radioactivity in samples of soil and sediment. The most cost-effective method is by high-resolution, low-level gamma spectrometry using high-purity germanium (HPGe) detectors (Murray *et al.* 1987; Wallbrink *et al.* 2002; Foster *et al.* 2007; Gilmore 2008; Appleby 2013). Radionuclides within the soil or sediment emit gamma photons at known energies and these interact with the germanium (cooled to liquid nitrogen temperature) in the detector. There are several alternatives for the detector configuration, which will determine the energy range over which they perform best. However, when dealing with small sample masses (e.g. lake bottom or fluvial suspended sediment samples), the choice of detector configuration is limited and a well detector, with a re-entrant volume for the sample, is commonly used. This geometry is highly efficient as the sample is 'surrounded' by the Ge in the detector. If a suitable detector is available, measurements of ^{137}Cs, ^{7}Be and ^{210}Pb can be made simultaneously.

^{210}Pb decays further to ^{210}Po, and several researchers have measured the alpha decay of ^{210}Po to estimate ^{210}Pb activities, because it provides more accurate results than the direct measurement of ^{210}Pb by gamma spectrometry (Joshi and Mudroch 1988; Zaborska *et al.* 2007; Kirchner 2011). However, these measurements are much more demanding in terms of laboratory pre-processing of samples. Furthermore, in order to date sedimentary sequences it is necessary to calculate the $^{210}Pb_{un}$ activity of a sample. To do this, the activity of ^{226}Ra is required, as this provides an estimate of the supported ^{210}Pb activity in the sample (see the section 'Dating sediment'), which cannot be measured directly by alpha decay. Some workers have assumed that sediment from the base of a core can be used to provide an estimate of the supported ^{210}Pb activity. However, Brenner *et al.* (2004) and others (e.g. Foster *et al.* 2006) have argued that ^{226}Ra activity should be measured on all samples, as changes in the source of the accumulating lake sediment are, for example, likely to change the activities of ^{226}Ra.

Samples are usually prepared for analysis by oven drying, disaggregating in a pestle and mortar and sieving to a pre-determined particle size (generally <2 mm). Since particle size commonly exerts an important influence on activity, it is important that it is taken into account and specified. Whereas sample holders may vary in volume and specification depending on the detector geometry, samples must be packed to a constant depth, density and volume so that results are directly comparable to the sample geometry used to calibrate the detection system. A special requirement for the analysis of $^{210}Pb_{un}$ is that samples need to be packed into a high-density gas-impermeable sample holder (e.g. PTFE) and carefully sealed in order to prevent ^{222}Rn gas escape. Once packed, samples are usually stored for a minimum of 21 days to allow the ^{214}Pb activities to equilibriate with ^{226}Ra (Appleby *et al.* 1986; Appleby 2001, 2008). The count times for sample analysis will vary depending on sample mass, geometry and the activities of individual radionuclides in a sample. These will be unique to individual samples and measurement systems, but the objective is normally to minimize the count time while achieving a counting error of $<5\% \pm 1$SD.

Several of the nuclides listed in Table 9.1 emit photons at more than one energy. Where possible, activities are measured at two energies for quality control purposes and the average activities and counting errors are reported. ^{235}U, measured at 186 keV, coincides with a weak ^{226}Ra gamma emission. Murray *et al.* (1987) suggested that the activity attributable to ^{226}Ra at this energy could be calculated from ^{214}Pb activities and subtracted from the total activity at this energy. Since ^{7}Be has a short half-life (~53 days), storage times need to be taken into account in reporting activities by correcting for decay between the final day of measurement and the date of sampling. Where field experiments are run or samples collected over several years, corrections will also be required for the decay of ^{137}Cs.

Those radionuclides listed in Table 9.1 which have either a complex fallout history (e.g. ^{137}Cs) or short half-lives (e.g. ^{7}Be and $^{210}Pb_{un}$) cannot be used in a longer term historical context for tracing, whereas those with half-lives of 10^3–10^9 years (^{235}U, ^{238}U, ^{40}K) can be assumed to be reasonably conservative and used in tracing studies spanning decades to centuries or longer.

Environmental magnetism

Studies of environmental magnetism make use of measurements of magnetic susceptibility and magnetic remanence (Thompson and Oldfield 1986; Walden *et al.* 1999; Dearing 2000; Oldfield 2007). There are several reasons for making

Table 9.2 Mineral magnetic properties used in the analysis of soils and sediments.

Property	Measured (M)/ derived (D)	Units
κ	M	Volume susceptibility
χ_{lf}	M	10^{-6} m^3 kg^{-1}
χ_{hf}	M	10^{-6} m^3 kg^{-1}
χ_{fd}*	D	10^{-9} m^3 kg^{-1}
$\chi_{fd\%}$†	D	%
$ARM_{(40\,\mu T)}$	M	10^{-3} A m^2 kg^{-1}
$IRM_{(0.25–1.0\,T)}$‡	M	10^{-3} A m^2 kg^{-1}
IRM_{loss} (24 h)	M	%
$IRM_{(-0.1\,T)}$	M	10^{-3} A m^2 kg^{-1}
χ_{arm}§	D	10^{-6} m^3 kg^{-1}
S ratio¶	D	Dimensionless
HIRM**	D	10^{-3} A m^2 kg^{-1}
MDF††	M	mT

* χ_{fd}: $[(\chi_{lf} - \chi_{hf})/m] \times 100$ (m = sample mass).
† $\chi_{fd\%}$: $[(\chi_{lf} - \chi_{hf})/\chi_{lf}] \times 100$.
‡At 1.0 T, samples are assumed to saturate and are referred to as saturated isothermal remanent magnetization (SIRM).
§ χ_{arm}: ARM × 3.14 × 10.
¶*S* ratio: $-1 \times (IRM_{-0.1T}/IRM_{1.0T})$.
**HIRM: $[IRM_{1.0T}/(1 - S\text{ ratio})]/2$.
††The median destructive field (MDF) of (S)IRM is the field at which a SIRM is demagnetized to 50% of its original value. These measures can help discriminate between MD magnetite and SD magnetite and or magnetite and hematite (Hatfield and Maher 2009; Maher *et al.* 2009).
Adapted from Higgitt *et al.* (1991), Walden *et al.* (1999), Foster *et al.* (2008), Hatfield and Maher (2009) and Maher *et al.* (2009).

these measurements. Pedologists use magnetic signatures in order to characterize the mineralogy and size of the magnetic grains present in soils, whereas geomorphologists use the measurements to characterize sediment sources or to provide points of time-synchronous correlation in sedimentary deposits (Walling *et al.* 1979; Caitcheon 1993, 1998; Verosub and Roberts 1995; Foster *et al.* 1998, 2008; Foster and Lees 2000; Walling 2005; Hatfield and Maher 2009; Maher *et al.* 2009; Boyle *et al.* 2010).

The following paragraphs provide a brief introduction to the measurement systems and data interpretation. A summary of the parameters used and the units in which they are measured is given in Table 9.2.

Magnetic susceptibility

Magnetic susceptibility measures the degree to which a sample can be magnetized while it is exposed to a weak magnetic field. The ratio of the magnetization (M) produced in a sample to the intensity of the magnetic field applied (H) defines the volume susceptibility, which is normally measured in the field and is designated by the symbol kappa (κ). When measured in the laboratory, it is reported on a mass-specific basis, referred to as mass-specific susceptibility, and designated by the symbol chi (χ). Susceptibility can be measured at different alternating frequencies, from which frequency-dependent susceptibility can be calculated on a concentration basis (χ_{fd}) or as a percentage ($\chi_{fd\%}$). κ and χ_{lf} measure the sum of all magnetic susceptibilities of a range of environmental materials including, from strongest to weakest, ferrimagnets (e.g. magnetite), anti-ferromagnets (e.g. haematite), magnetically weak paramagnetic materials (e.g. pyrite and biotite) and diamagnetic material (e.g. water, chalk and organic matter). Diamagnetic materials produce low negative values of susceptibility.

χ_{lf} ranges from –0.01 to ~1000 × 10^{-6} m^3 kg^{-1} in environmental samples, with high values associated with ferrimagnetic minerals, basic/ultrabasic rocks and soils that have been exposed to fire [a property that can make magnetic measurements useful for determining contributions from eroded topsoil following forest fires (e.g. Oldfield 1991; Blake *et al.* 2006; Oldfield and Crowther 2007)]. Values for sedimentary rocks range from ~0.001 to 0.1 × 10^{-6} m^3 kg^{-1} whereas metamorphic and igneous rocks produce intermediate values (Dearing 1999, 2000).

Frequency-dependent susceptibility is used to detect the presence of very small magnetic grains (~0.02 μm in diameter), which exhibit viscous, time-dependent behaviour. These grains are often produced in soils or lake sediments as a result of weathering and pedogenesis or by burning (Dearing *et al.* 1996; Oldfield 1994; Blake *et al.* 2006; Oldfield and Crowther 2007). Further details of the theory and the measurements, along with tables of typical magnetic susceptibilities for a range of environmental materials, are provided by Thompson and Oldfield (1986) and Dearing (1999, 2000).

Magnetic remanence

Remanence properties are measured after the sample has been magnetized and removed from the magnetic field. The strength of the acquired magnetic remanence is measured in a spinner magnetometer. Several methods exist for magnetizing samples, but anhysteretic remanent magnetization (ARM) and isothermal remanent magnetization (IRM) are often measured routinely in the laboratory (Thompson and Oldfield 1986; Maher 1988; Walden *et al.* 1999; Maher *et al.* 2009; Hatfield and Maher 2009). Once the sample acquires no further remanence with an increase in the applied field, this is referred to as saturated remanence (SIRM).

Following saturation, samples may be left for 24 hours and re-measured in order to estimate the quantity of viscous grains that lose their remanence slowly over time. IRM_{loss} is usually expressed as a percentage of that measured at saturation (Higgitt *et al.* 1991; Foster *et al.* 2008).

Finally, samples are often magnetized after saturation in a reverse field (commonly −100 mT) to establish the ease with which the remanence can be reversed after saturation. Values can be expressed on a concentration basis (HIRM) but these values can also be used to calculate the *S* ratio (see Table 9.2).

Hatfield and Maher (2009) and Maher *et al.* (2009) reported measuring a demagnetization parameter, the median destructive field (MDF), which measures the field required to reduce the SIRM to 50% of its original value. This can be achieved by demagnetizing the sample at slowly increasing fields in an alternating field demagnetizer fitted with a tumbling rotation arm (Walden 1999).

Where samples contain a high proportion of weakly diamagnetic organic matter, susceptibility and remanence concentration parameters are usually corrected by determining the organic matter content by loss on ignition. In some cases, small secondary minerals (including bacterial magnetites) or magnetic mineral coatings on larger grains are removed by acid pre-treatment of the samples before measurement. This leaves magnetic inclusions, protected within host silicate grains, for example, as the basis of the measured magnetic signature of a sample (e.g. Hounslow and Morton 2004; Maher *et al.* 2009).

Sediment geochemistry

Modern developments in analytical chemistry have provided a range of techniques for rapidly determining the concentration of suites of geochemical elements and stable isotopes in fine sediment. These fall broadly into two groups of signatures; organic and inorganic.

Organic signatures include C, N, the C : N ratio and stable carbon and nitrogen isotope signatures (e.g. O'Malley *et al.* 1996; Kaushel and Binsford 1999; Papanicolaou *et al.* 2003; Turnbull *et al.* 2008; Mahapatra *et al.* 2011). Inorganic signatures include a wide range of stable geochemical elements (such as Al, As, Ba, Bi, Cd, Ce, Co, Cr, Cs, Cu, Dy, Er, Eu, Fe, Ga, Gd, Ge, Hf, Ho, In, K, La, Li, Mg, Mn, Mo, Na, Nd, Ni, P, Pb, Pd, Pr, Rb, Sb, Sc, Sm, Sn, Sr, Tb, Ti, Tl, U, V, Y, Yb, Zn and Zr) (e.g. Collins and Walling 2004; Foster *et al.* 2007; Collins *et al.* 2010a). A number of stable isotopes and ratios, such as $^{87}Sr : ^{86}Sr$ and $^{206}Pb : ^{207}Pb$, have been used in studies determining long-term sediment source changes in river catchments and for assessing the relative significance of atmospheric pollution in contributing to Pb contamination in sediments (e.g. Talbot *et al.* 2000; Krom *et al.* 2002; Graham *et al.* 2006; Vane *et al.* 2011; Mighall *et al.* 2014). Heavy metal speciation and stable isotope signatures have also recently been used to distinguish geogenic from urban sediment sources (e.g. Le Pape *et al.* 2014; Thapalia *et al.* 2015; Wiederhold 2015).

A significant consideration in selecting the most appropriate suite of elements for analysis is that signatures are conservative through time. For tracing actively transported sediments, most of these elements can be considered to be reasonably conservative although, like mineral magnetic and radionuclide signatures, they will require screening and correction for particle size effects (see the section 'Sediment source fingerprinting'), although several recent papers have questioned the validity and benefits of both organic matter and particle size corrections (e.g. Smith and Blake 2014; Pulley *et al.* 2015a). For sedimentary deposits, these assumptions do not always hold, as several of the signatures cannot be assumed to have remained constant through time. This is especially true of those elements that have increased in concentration in topsoil due to atmospheric pollution (e.g. the heavy metals Cu, Pb, Ni and Zn) or where they have increased in concentration in soils due to their use as fertilizers (e.g. Foster and Charlesworth 1996; Foster and Lees 2000; Withers *et al.* 2001). Boyle *et al.* (2010) suggest that secondary ferromagnetic minerals growing in soils may be slow enough for the magnetic signatures to be used over time-scales of decades to centuries, but probably not for millennia or longer periods of geological time. This presupposes, however, that the signatures are not affected by post-depositional dissolution or diagenesis or by the in-growth of bacterial magnetite (e.g. Foster *et al.* 2008; Pulley *et al.* 2015b).

Analytically, the inorganic elements can be analysed by non-destructive methods (e.g. X-ray fluorescence or electron microprobe) or destructive methods [acid digestion followed by atomic absorption spectrometry (AAS) or inductively coupled plasma atomic emission spectrometry (ICP-AES)]. Specialist stable isotope laboratories are required to analyse a wide range of potentially useful isotope concentrations and ratios. A detailed discussion of digestion and analytical methods is beyond the scope of this chapter and readers are referred to Hassan *et al.* (2007), Murphy and Morrison (2007) and Baskaran (2011).

9.3 Applications

Dating sediment

Both ^{137}Cs and ^{210}Pb can be used to date sedimentary deposits over the last ~100 years. The use of both approaches in combination has the advantage that the results are based on different assumptions and ^{137}Cs can be used to test the validity of the depth–age curve produced by ^{210}Pb dating. The following sections outline the principles that underpin the two techniques.

Dating using ^{210}Pb

A number of different models have been developed to calculate the age of depositional layers in a lake sediment deposit from the changing activity of $^{210}Pb_{un}$ in a profile. In all cases, $^{210}Pb_{un}$ is the amount of the ^{210}Pb isotope that is in excess of the background ^{210}Pb produced in accumulating sediments by ^{226}Ra decay. $^{210}Pb_{un}$ reaches the lake either directly via atmospheric fallout to the lake surface or indirectly as the $^{210}Pb_{un}$ accumulating in catchment soils is eroded and transported to the point of deposition. Early models, e.g. the constant flux–constant sedimentation (CFCS) model of Robbins (1978) or the constant initial concentration (CIC) model of Appleby and Oldfield (1978), used to interpret the $^{210}Pb_{un}$ depth distribution, made very simple assumptions about constant sedimentation rates through time and the delivery of $^{210}Pb_{un}$ to the point of deposition. However, geomorphologists are frequently interested in disturbed systems, where sedimentation rates vary rapidly and these models fail to account for such variability. Models that can account for fluctuations in sedimentation rate include the Appleby and Oldfield (1978) constant rate of supply (CRS) model and a variant of this model, the composite-CRS (C-CRS) model, which can be applied when specific depths in the sediment column can be dated by other means (e.g. the known age of a reservoir or other time synchronous marker such as pollution history) (Appleby 2001, 2008). Detailed explanations of these methods are provided by Appleby (2001, 2008) and Du and Walling (2012), while a personal reflection on the development of the ^{210}Pb dating methods is given by Appleby (2013).

The models above were developed specifically for use in marine and lake sediments. He and Walling (1996), however, describe a ^{210}Pb dating model; the constant initial concentration and constant sedimentation rate (CICCS) model, that was specifically developed for use in floodplains. The model only uses the total $^{210}Pb_{un}$ inventory for a whole core and produces an estimate of the average accumulation rate for the last 100 years of sedimentation. Further detailed discussion of the ^{210}Pb dating models that would be applicable to floodplains and colluvial deposits is provided by Du and Walling 2012.

Dating floodplain or colluvial accumulation rates using ^{210}Pb is more complex than dating marine and lake sediment sequences, as there is a higher probability that some escape of ^{222}Rn gas will occur through the pore spaces or pore water in the sediment and be released to the atmosphere. Indeed, escape of ^{222}Rn from soil is the mechanism by which $^{210}Pb_{un}$ is delivered to lake and ocean surfaces and provides the basis for ^{210}Pb dating. It is therefore unrealistic to assume that the ^{214}Pb activity will be in equilibrium with the ^{226}Ra activity, which is generally assumed to be the case when using gamma spectrometry to determine ^{226}Ra activity by measuring ^{214}Pb. This loss is referred to in the literature as the emanation coefficient and its value appears to lie somewhere between 0.07 and 0.7 depending on soil type (Du and Walling 2012; Table 9.2), with most estimates for soils ranging between 0.2 and 0.3. Evidence of the impact of ^{222}Rn loss on the estimation of $^{210}Pb_{un}$ is presented in Fig. 9.1, for a floodplain core collected from the River Axe in Devon, England, by assuming a range of emanation coefficients. Setting the coefficient to ~0.3 (Fig. 9.1e) eliminates

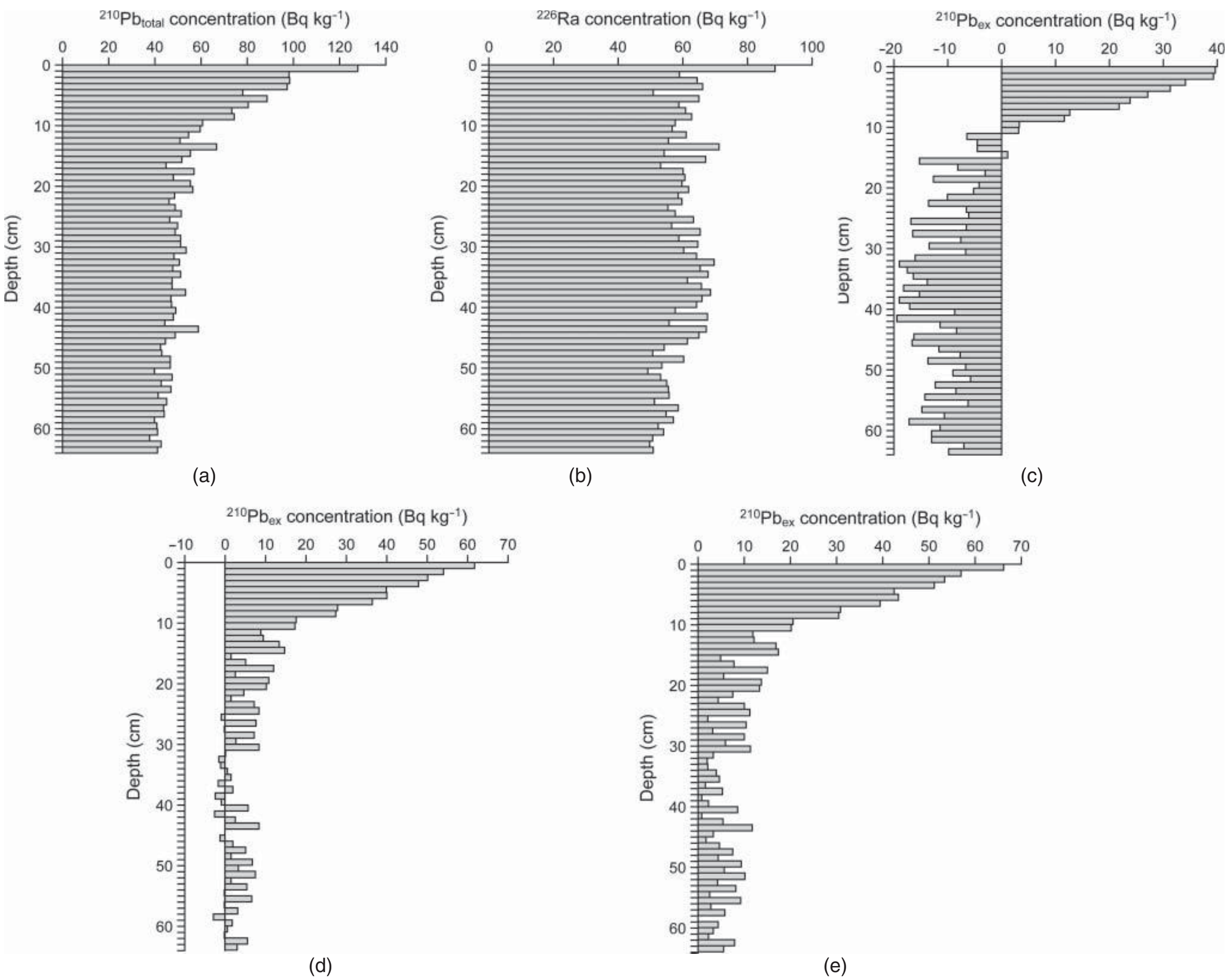


Figure 9.1 The $^{210}Pb_{total}$ (a) and ^{226}Ra (b) profiles for a sediment core collected from the floodplain of the River Axe at Nunford Bridge, Devon, UK. The vertical distributions of $^{210}Pb_{un}$ obtained when applying an emanation coefficient of 0, 0.25 and 0.30 are shown for the same core in (c), (d) and (e), respectively. Source: Du and Walling, 2012. Reproduced with permission of Elsevier.

the 'apparent' negative values in the $^{210}Pb_{un}$ profile. In the same study, Du and Walling (2012) tested several of the ^{210}Pb dating models originally developed for lakes and concluded that the CICCS and C-CRS models appeared generally to give the best results for floodplain sites in the United Kingdom.

Dating using ^{137}Cs

As indicated earlier in the section 'Gamma-emitting radionuclides', ^{137}Cs is an atmospheric fallout radionuclide derived both from atmospheric nuclear weapons testing in the late 1950s and early 1960s and, over large parts of Europe and adjacent regions, from fallout from the Chernobyl nuclear accident in 1986. The latter produced more spatially variable fallout than the weapons fallout, as it followed a complex pattern of movement at relatively low levels in the atmosphere across northern and southern Europe and produced extremely high fallout in areas where rain occurred as the cloud passed overhead (Walling *et al.* 1989; Foster 1991). In contrast to $^{210}Pb_{un}$, where the chronology is derived from the decrease in $^{210}Pb_{un}$ activity with depth, the use of ^{137}Cs relies on the existence of chronological marker horizons that reflect the temporal record of fallout. These markers relate to the onset and peak of bomb fallout and the occurrence of Chernobyl fallout in 1986. The timing of the bomb fallout marker horizons differs slightly between the northern and southern hemispheres. Whereas significant fallout was first recorded around 1954 and peaked in 1963 in the northern hemisphere, it was not recorded in significant amounts until about 1956 in the southern hemisphere and peaked around 1965 (Rowntree and Foster 2012). There are no reports of an input from the 1986 Chernobyl nuclear accident transferring ^{137}Cs to the southern hemisphere.

Unlike ^{210}Pb, ^{137}Cs does not provide a continuous sequence of dates in a sedimentary deposit. In this respect, it is analogous to other age-equivalence dating methods such as tephra chronologies (Walker 2005) and dating based on the fallout of spherical carbonaceous particles derived from early fossil fuel combustion (e.g. Rose *et al.* 1995; Pittam *et al.* 2009). As fallout levels were low in the southern hemisphere, it is now unlikely that the first occurrence of fallout will be detectable in sedimentary sequences, because activities are unlikely to exceed limits of detection on gamma spectrometry systems. However, a significant increase in ^{137}Cs fallout occurred in ~1958 (Foster and Rowntree 2012; Foster *et al.* 2012), and it is likely that this level will still be detectable in sedimentary deposits. The possibility that the level above which ^{137}Cs is first found in a sediment profile may be influenced by downward diffusion or migration of the radionuclide must, however, also be considered. Such downward displacement is more likely to occur in floodplain sediments than in lake or marine sediment, where the occurrence of bioturbation and related processes is likely to be restricted, and for this reason caution should be exercised in using the first appearance of ^{137}Cs as a time marker in floodplain sediments. More detailed descriptions and interpretation of patterns of ^{137}Cs fallout as recorded in lake sediments are provided by Walling and He (1992), He *et al.* (1996) and Zhang and Walling (2005).

Limitations to ^{210}Pb and ^{137}Cs dating

As indicated above, radionuclides deposited and assimilated into sediment sequences may show some evidence of downward migration and diffusion over time as a result of either biological activity or chemical diffusion. This problem is likely to be more important for terrestrial deposits than for lake and marine sediments. Equally, it is likely to prove more significant for ^{137}Cs profiles than for $^{210}Pb_{un}$, since ^{137}Cs has a known atmospheric fallout history which is expected to be reflected by a similar pattern in its depth distribution with sedimentary deposits.

Although relatively uncommon, there are published reports of significant re-mobilization and release of radionuclides from lake sediments. For example, Benoit and Hemond (1991) measured ^{210}Pb, ^{210}Po and ancillary geochemical parameters in the sediments and pore waters of a lake with seasonally anoxic bottom waters and found significant release of radionuclides to the water column. Their analysis demonstrated that solid-phase ^{210}Pb profiles did not match the expected input history, suggesting that the radionuclide may be undergoing redistribution and loss. Brenner *et al.* (2004) reported examples from lakes in Florida where core inventories were augmented by inputs of groundwater containing significant amounts of dissolved ^{226}Ra. This ^{226}Ra is adsorbed by recent sediments and complicates accurate estimation of supported ^{210}Pb activity, and confounds calculation of the $^{210}Pb_{un}$ activity required by the dating models outlined above.

Foster *et al.* (2006) reported the failure of ^{137}Cs to date sediment sequences in coastal lagoons in southwest England. Here it was demonstrated that remobilization of ^{137}Cs most probably occurred as a result of seawater intrusion through permeable gravel and peat layers beneath the lagoon barriers during very high tides, when sea levels were much higher than the lagoon levels. Several studies have suggested that ^{137}Cs is more mobile than ^{210}Pb and that a number of factors could be responsible for re-mobilization, including reduced oxygen levels, high salinity (displacement with Na^+) and the presence of NH_4^+ and H^+ (Longmore *et al.* 1983; Appleby 2001; Foster *et al.* 2006).

Additional problems arise in environments where the sources of sediment delivered from a catchment to a point of deposition have changed over time, since this can influence the $^{210}Pb_{un}$ and ^{137}Cs activity of catchment-derived sediment and, thus, the depth distribution of the radionuclide. Such changes can be particularly important when attempting to match the ^{137}Cs depth distribution to the fallout record. Examples of a change from surface to sub-surface domination of sediment sources during the period of maximum atmospheric ^{137}Cs fallout, resulting in a marked reduction in ^{137}Cs activity unrelated to the fallout record, were reported in two palaeoenvironmental reconstructions using sediments deposited in South African farm dams by Rowntree and Foster (2012) and Foster and Rowntree (2012).

Documenting soil and sediment redistribution

The use of fallout radionuclides to document rates and patterns of recent soil and sediment redistribution within the landscape must be seen as representing a major and timely advance in the geomorphologist's capacity to document contemporary processes. It has helped to address the need for the spatially distributed information necessary to establish catchment sediment budgets and to develop and validate catchment sediment yield models. Increasingly, such data also have important practical applications in developing catchment sediment management strategies (Walling and Collins 2008). The approach is founded on the existence of a number of radionuclides, both natural and man-made, that reach the land surface as fallout, primarily during rainfall events and are rapidly and strongly fixed by the surface soil or sediment. The subsequent redistribution of these radionuclides within a catchment or river system is a direct reflection of the movement of the soil or sediment particles to which the radionuclides are attached. By studying the post-fallout redistribution and fate of these fallout radionuclides, it is possible to obtain essentially unique information on soil and sediment redistribution and, therefore, on erosion and deposition rates. Fallout radionuclides with different half-lives can be employed to provide information relating to different time-scales.

The fallout radionuclide most widely used for this purpose is ^{137}Cs. As indicated in the earlier section 'Gamma-emitting radionuclides', ^{137}Cs is a man-made radionuclide that was produced by the testing of thermonuclear weapons in the 1950s and early 1960s, with the Chernobyl accident providing further fallout inputs to adjacent regions in 1986. Caesium-137 has a half-life of 30.2 years and much of the original fallout will therefore still remain within the upper horizons of the soils and sediments of a catchment. By investigating the current distribution of the radionuclide in the landscape, it is possible to obtain information on the net effect of soil and sediment redistribution processes operating over the past ~50 years (Walling and Quine 1991; Zapata 2002) or, where attention focuses on Chernobyl fallout, over the past ~30 years (Golosov 2002).

When sampling the soils and sediments in a study area, attention is commonly directed to both the areal activity density or inventory (Bq m^{-2}), which represents the total amount of radionuclide contained within the soil per unit surface area and the depth distribution. The latter is generally expressed in terms of the variation of mass activity density (Bq kg^{-1}), usually referred to as the activity, in relation to depth, expressed as either a linear depth (m) or a mass depth (kg m^{-2}). Some typical examples of the depth distribution of ^{137}Cs in undisturbed and cultivated soils are presented in Fig. 9.2. In an undisturbed soil, the depth distribution is commonly exponential in form and most of the ^{137}Cs is found in the upper ~12 cm. In contrast, in a cultivated soil, the ^{137}Cs will be mixed into the soil by the tillage and the ^{137}Cs activity will be fairly uniform within the plough layer. Removal of soil from the surface by erosion will result in a reduction of the inventory, but the magnitude of the reduction

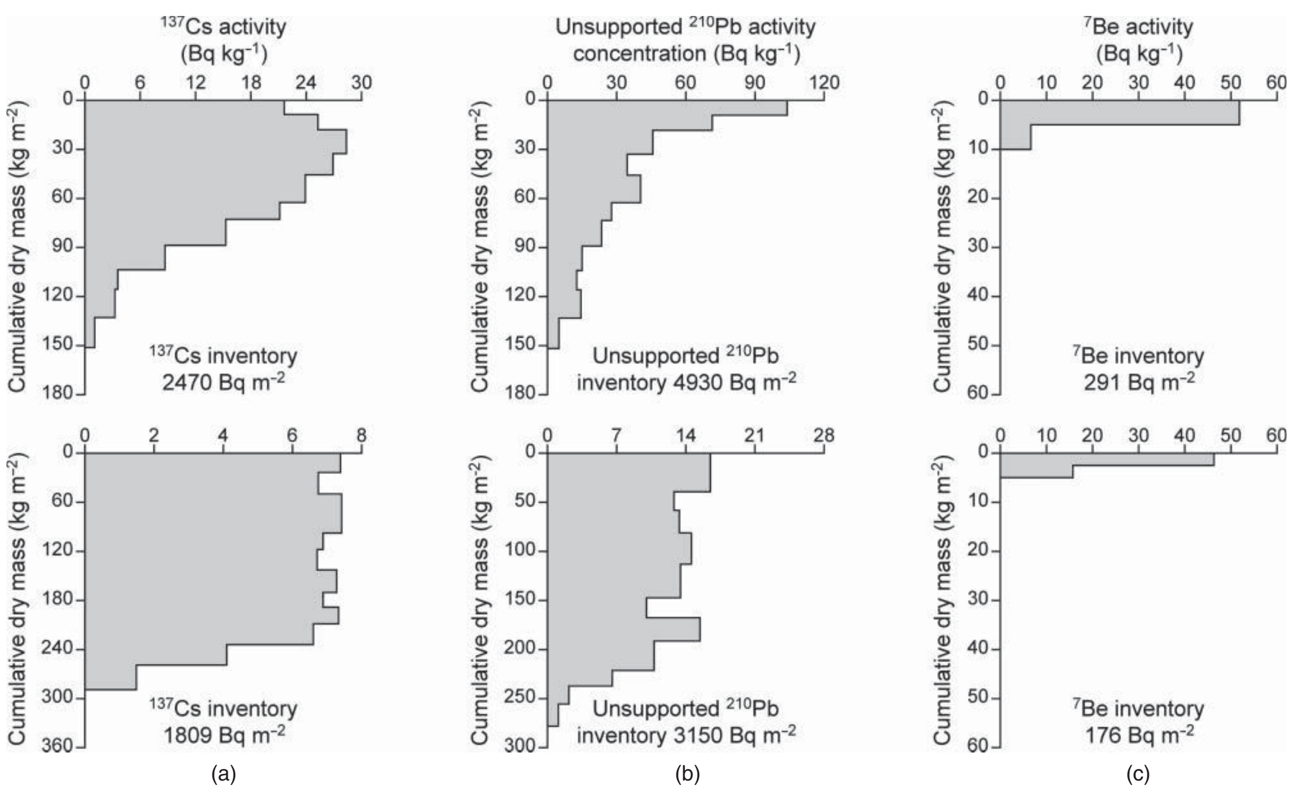


Figure 9.2 Typical depth distributions of (a) ^{137}Cs, (b) $^{210}Pb_{un}$ and (c) ^{7}Be in undisturbed pasture (above) and cultivated soils (below) in Devon, UK.

and the effect on the depth distribution will vary between an uncultivated and a cultivated soil. In the former case, the depth distribution will be truncated by removal of the surface layer. In a cultivated soil, continued cultivation and associated mixing will maintain the uniform activity within the plough depth by incorporating soil from beneath the original plough depth, but the activity of the plough layer will decline as new soil is added. Deposition causes addition of soil or sediment to the surface, causing an increase in the inventory and an upward extension of the depth distribution.

In order to reduce the number of samples requiring analysis, the use of ^{137}Cs measurements to document soil redistribution rates commonly relies primarily on bulk cores, which provide values of the total ^{137}Cs inventory for the sampling points. These values are compared with the local reference inventory, which represents the inventory associated with a flat, undisturbed site where neither erosion nor deposition has occurred (Pennock and Appleby 2002). Reduced inventory values will denote an eroding point, whereas an increased inventory will reflect deposition. Mean soil redistribution rates over the past ~50 years, since the main period of fallout, can be estimated from the degree of departure of the measured inventory from the reference value, using a range of conversion models (e.g. Walling and He 1999a; Walling *et al.* 2002a; Li *et al.* 2009). A useful discussion of sampling methods is provided by Loughran *et al.* (2002). Careful thought also needs to be given to sampling schemes, in order to provide representative results (Pennock and Appleby 2002), and both transects and grid sampling schemes are often used. Figure 9.3(a) and Table 9.3 present the results of a study involving the use of ^{137}Cs measurements to obtain information on the rates and pattern of medium-term soil redistribution within a 6.7 ha field in Devon (Walling *et al.* 1999a). In this case, 140 bulk cores were collected at the intersections of a 20 m × 20 m grid. Such data are essentially impossible to obtain using other approaches.

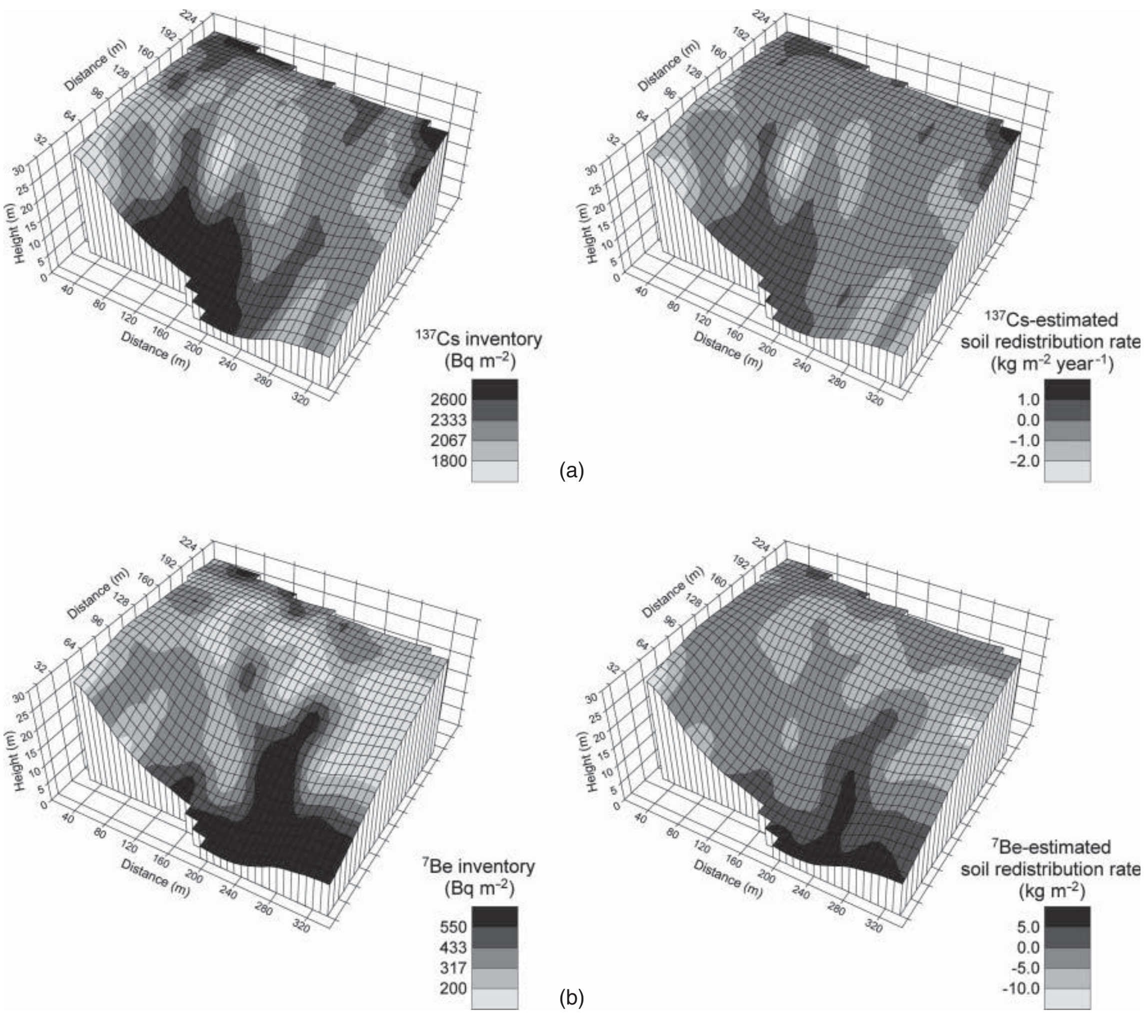


Figure 9.3 The spatial pattern of (a) ^{137}Cs and (b) ^{7}Be inventories within a field near Crediton, Devon, UK and the estimates of soil redistribution rates derived from these measurements.

Table 9.3 A comparison of the rates of soil redistribution within the study field shown in Fig. 9.2 estimated from the ^{137}Cs and ^{7}Be measurements.

Parameter	^{137}Cs (kg m^{-2} year^{-1})	^{7}Be (kg m^{-2})
Range	−4.5 to +2.0	−11.9 to +9.8
Mean erosion rate in eroding area	−1.1	−5.3
Mean deposition rate in deposition area	0.69	4.0
Net soil loss	−0.48	−2.5
Field sediment delivery ratio	0.83	0.80

Caesium-137 measurements have also been successfully used to estimate overbank sedimentation rates on river floodplains over the past ∼50 years (Walling and He 1993, 1997; Terry *et al.* 2002; Ritchie *et al.* 2004; Du and Walling 2012). In most studies, this has involved analysing sectioned cores from a floodplain and using the position of the peak ^{137}Cs activity in the core to identify the level of the floodplain surface at the time of peak fallout in the early 1960s. However, because a considerable number of ^{137}Cs measurements are needed to define the depth profile, the number of points for which sedimentation rates can be determined is necessarily limited. In order to increase the density of sampling points, procedures for estimating the sedimentation rate from measurements of the total inventory of a bulk core and comparing this value with the local reference inventory have also been developed (Walling and He 1997). Figure 9.4(a) presents an example of the application of ^{137}Cs measurements to document medium-term overbank sedimentation rates on the floodplain of the River Severn, near Buildwas, Shropshire, England. In this case, 124 bulk cores were collected from the study reach at the intersections of a 25 m × 25 m grid.

Although most studies employing fallout radionuclides to document rates and patterns of soil and sediment redistribution in the landscape have been based on ^{137}Cs, both ^{210}Pb$_{un}$ and ^{7}Be have also been used in a similar manner (Mabit *et al.* 2008; IAEA 2014). Typical depth distributions of these radionuclides in uncultivated and cultivated soil are presented in Fig. 9.2. The use of ^{210}Pb$_{un}$ to document soil and sediment redistribution within the landscape employs similar assumptions and procedures to those associated with ^{137}Cs. Mabit *et al.* (2014) provide a useful overview of the application of this radionuclide, Walling and He (1999b) discuss its use in soil erosion studies and He and Walling (1996) and Du and Walling (2012) provide examples of its application for estimating rates of overbank sedimentation on river floodplains. The half-life of ^{210}Pb is 22.3 years, which is similar to that of ^{137}Cs. However, because ^{210}Pb is a natural geogenic radionuclide, ^{210}Pb$_{un}$ fallout can be viewed as having been essentially constant through time and the inventory at a sampling point will reflect fallout receipt and subsequent redistribution and decay over the past ∼100 years. Measurements of ^{210}Pb$_{un}$ activity can therefore provide information on soil and sediment redistribution rates over the past ∼100 years and use of both ^{137}Cs and ^{210}Pb$_{un}$ in combination can provide additional information on the erosional or depositional behaviour of a study area (He and Walling 1996; Walling *et al.* 2003b). Figure 9.4 compares the estimates of overbank sedimentation rates on the floodplain of the River Severn at Buildwas provided by ^{210}Pb$_{un}$ measurements (b) with those provided by ^{137}Cs measurements (a). The similarity of the two sets of values suggests that overbank sedimentation rates along this reach have remained relatively stable over the past ∼100 years.

In contrast to ^{137}Cs and ^{210}Pb, ^{7}Be has a very short half-life of only 53 days, and it can be used to provide information on soil and sediment redistribution rates associated with individual events or short periods of heavy rainfall extending over a few weeks (e.g. Walling *et al.* 1999a; Blake *et al.* 2002; Wilson *et al.* 2003; Schuller *et al.* 2006; Sepulveda *et al.* 2007). The principles involved are similar to those for ^{137}Cs and ^{210}Pb$_{un}$, but for most applications it is important to ensure that the period investigated conforms to a number of requirements, in order to avoid carry-over effects from previous periods of heavy rainfall, which could influence the magnitude and spatial distribution of ^{7}Be inventories across the study area. This can limit the potential of the approach. However, Walling *et al.* (2009) and Porto *et al.* (2014) have recently described procedures for employing ^{7}Be measurements that largely overcome this constraint and make the approach more generally applicable over longer periods. Figure 9.3(b) and Table 9.3 illustrate the use of ^{7}Be measurements to obtain information on the soil redistribution in this 6.7 ha field caused by a period of heavy rainfall (69 mm in 7 days) during the winter of early 1998, when the field had been left bare and compacted after the fodder maize harvest (Blake *et al.* 1999; Walling *et al.* 1999a). The erosion associated with this single event in 1998, estimated using the ^{7}Be measurements, is considerably greater than the mean annual erosion rate for the field estimated using ^{137}Cs measurements.

Caesium-137 has now been successfully used in many areas of the world to obtain hitherto essentially unavailable information on medium-term rates of soil and sediment redistribution (Ritchie and Ritchie 2008). Its value as a tracer has been promoted by the International Atomic Energy Agency (IAEA) (Zapata 2002). ^{210}Pb$_{un}$ and ^{7}Be have been less widely used to date, but their use is expanding (IAEA 2014). Key advantages associated with the use of fallout radionuclides include the ability to obtain retrospective information on medium-term soil redistribution rates, the need for only a single sampling campaign, the provision of spatially distributed information relating to the individual sampling points and the ability to collect information from the natural landscape, without the need to install plots or to otherwise constrain the location of the measuring points. Most applications of fallout radionuclides to date have involved relatively small areas, since this permits the collection of sufficient samples to obtain representative information on the spatial patterns of soil and sediment redistribution involved. There is a need for further work to establish procedures for using the approach to obtain information from

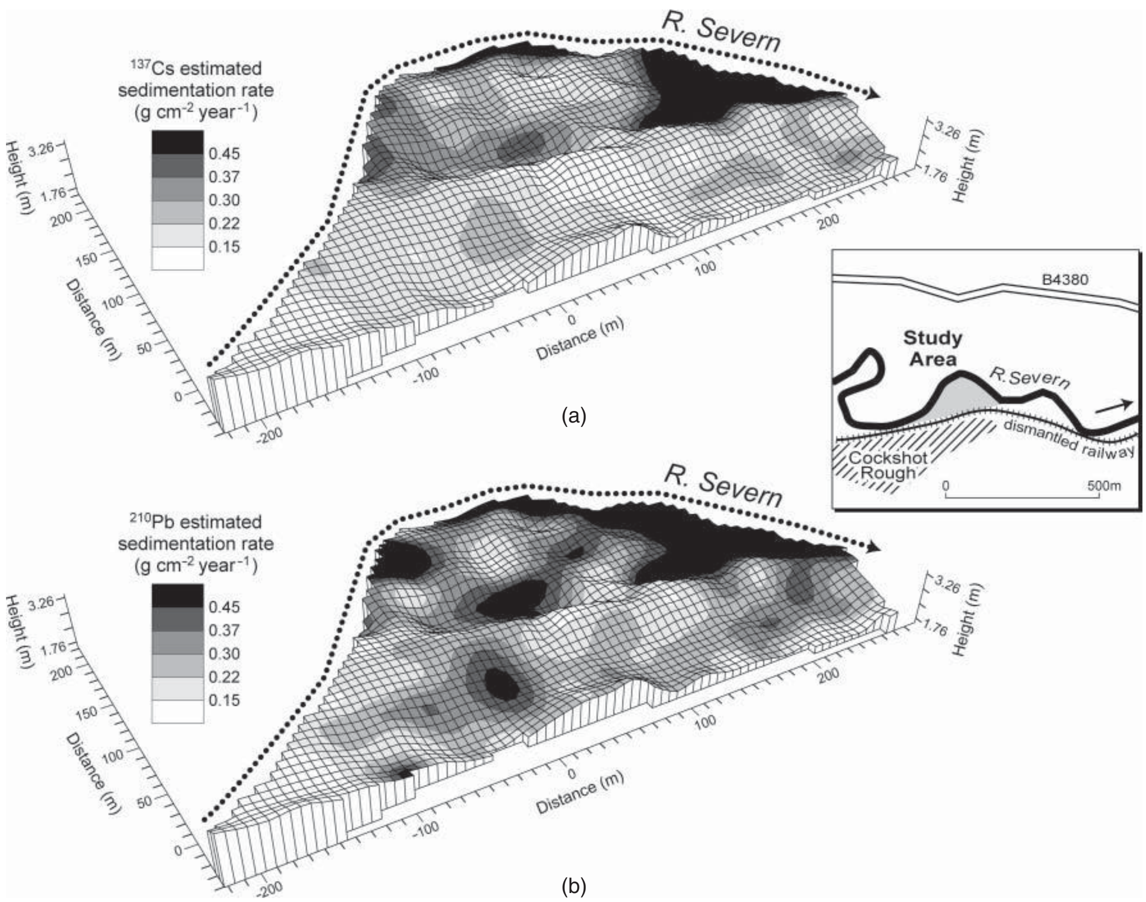


Figure 9.4 The spatial pattern of overbank sedimentation rates within a short reach of the River Severn near Buildwas, Shropshire, UK derived from (a) ^{137}Cs and (b) $^{210}Pb_{un}$ measurements made on floodplain cores.

larger areas without a major increase in the number of samples that need to be collected and analysed. The studies reported by Mabit *et al.* (2007), Porto *et al.* (2010), Walling and Zhang (2010), Chappell *et al.* (2011) and Walling *et al.* (2014) demonstrate progress in this direction. In most applications, ^{137}Cs and $^{210}Pb_{un}$ are used to provide estimates of medium- or longer term average soil redistribution rates (i.e. ~50 years and up to 100 years, respectively). However, with increasing interest in global change, attention has been directed to the potential for using fallout radionuclides to document changing erosion rates. In the case of ^{137}Cs, Schuller *et al.* (2007) report the use of ^{137}Cs measurements to document the change in soil redistribution rates associated with a switch from conventional to no-till management system in an area of cereal cultivation in south-central Chile and Porto *et al.* (2014) demonstrate the potential for using a re-sampling technique to provide estimates of soil redistribution rates for two different periods. In the case of $^{210}Pb_{un}$, its continuous fallout and its half-life of 22.3 years mean that inventories will be preferentially influenced by recent soil redistribution and potential exists to compare soil redistribution rates estimated using both radionuclides to identify recent changes (Porto *et al.* 2013).

Sediment source fingerprinting

Information on the source of the suspended sediment transported by a stream or river can be required for many purposes. The sources could be classified according to spatial location (e.g. sub-catchments or areas underlain by different soil or rock types) or source type (e.g. channel erosion or sheet and rill erosion from cultivated areas or areas under pasture or rangeland). Such information could, for example, be required when constructing a sediment budget for a catchment, attempting to interpret the measured sediment yield from a catchment in terms of erosion rates and landscape evolution and developing and validating catchment sediment yield models. Recent concern regarding the role of fine sediment in degrading aquatic ecosystems and habitats has highlighted the need to control sediment mobilization and transfer and information on sediment source is a key requirement for cost-effective targeting of control measures.

Collins and Walling (2004) have reviewed the various approaches that can be used to obtain information on sediment source. They distinguished both *indirect* and *direct* approaches to determining the relative contribution of different potential sources within a catchment or river basin. The former use techniques such as erosion pins or remote sensing to estimate the erosion rates associated with different processes or to identify erosion 'hotspots' and thereby infer the relative importance of different sources. The latter focuses on the sediment transported by the stream or river and uses 'fingerprinting' techniques to determine its source more directly. In this case, potential source materials are discriminated using their physical, mineral magnetic, geochemical and other properties. By comparing the properties of the transported sediment (the target) with those of the potential sources, it is possible to establish their relative contribution to the suspended sediment load. Because of the many uncertainties associated with indirect approaches, sediment source tracing or fingerprinting is generally seen as providing the most effective and reliable approach and it is being increasingly used to support the design and implementation of catchment sediment management programmes (Gellis and Walling 2011; Gellis and Munkundan 2013). The key components of sediment source fingerprinting are described below.

Walling (2013) has reviewed the history of the source fingerprinting approach, which can be traced back to the 1970s and the work of researchers such as Klages and Hsieh (1975), Wall and Wilding (1976) and Walling *et al.* (1979). In these studies, the sources were loosely defined and the assessment of their relative contribution was essentially qualitative. Since this early work, most studies have focused on discriminating source types, rather than spatial sources. Information on the latter could potentially be obtained by measuring and comparing the sediment loads of individual tributaries, but information on source type is difficult to obtain using other approaches. Subsequent refinement of the approach has involved a number of important developments that can also be seen as key elements of the source fingerprinting technique. These relate to the following:

1. the fingerprint properties employed;
2. the use of statistical tests to identify the most effective fingerprint properties;
3. the use of numerical mixing (or unmixing) models to permit quantitative assessment of the relative contribution of different potential sources;
4. testing for conservative behaviour of the fingerprint properties employed and taking account of enrichment/depletion effects;
5. taking account of uncertainty in the results;
6. application of the approach to an increased range of 'targets';
7. extension of the approach to incorporate a temporal dimension.

All these aspects are considered further below.

Fingerprint properties and composite fingerprints

Although some early studies made use of a single sediment property as a fingerprint, subsequent work has emphasized that a composite fingerprint, comprising several properties, will provide better discrimination between potential sources and therefore more reliable results. A wide range of soil and sediment properties have now been successfully used as fingerprints. In addition to mineralogy, mineral magnetic and geochemistry measurements (e.g. Walden *et al.* 1999; Slattery *et al.* 2000; Pulley *et al.* 2015a), colour and spectral reflectance (Grimshaw and Lewin 1980; Poulenard *et al.* 2009; Martínez-Carreras *et al.* 2010a, 2010b), isotopic signatures (Douglas *et al.* 1995, 2003; Fox and Papanicolaou 2007), fallout radionuclides (Walling and Woodward 1992; Olley *et al.* 1993; He and Owens 1995; Wallbrink *et al.* 1998, 1999), plant pollen (Brown 1985) and compound-specific stable isotopes (Gibbs 2008) have also been shown to provide effective fingerprints. As the range of fingerprint properties that have been successfully used for source discrimination has increased, so too has the range of potential sources that can be considered. Unmetalled roads and tracks (Gruszowski *et al.* 2003; Motha *et al.* 2004; Collins *et al.* 2010b) and damaged road verges (Collins *et al.* 2010c) have, for example, been successfully discriminated as potential sources and compound-specific stable isotopes offer the potential to distinguish sediment mobilized from areas under different crop or vegetation types (Gibbs 2008). Isotopic signatures have also been successfully used to discriminate sediment-associated heavy metals derived from urban and rural areas (Thapalia *et al.* 2015).

The vast range of fingerprint properties that has now been used in different studies could be seen as a problem, in terms of developing standardized procedures. In most studies, the fingerprint properties to be used are identified empirically, by analysing a range of properties and using statistical procedures to test for discrimination potential and to select the final composite fingerprint that affords maximum discrimination. Few attempts have as yet been made to develop general guidance for selecting appropriate fingerprint properties.

Statistical testing of source discrimination

As indicated above, the set of sediment properties to be included in a composite fingerprint is frequently selected empirically. The application of statistical tests such as the Mann–Whitney U-test (Carter *et al.* 2003; Porto *et al.* 2005), the Kruskal–Wallis H-test (Collins *et al.* 1998, 2001; Walling *et al.* 1999b), the Wilcoxon rank-sum test (Juracek and Ziegler 2009) and the Tukey test (Motha *et al.* 2003) to identify properties providing good discrimination and the use of discriminant function analysis and other classification techniques to select optimum combinations of those properties (Walling and Woodward 1995; Collins *et al.* 1998) have greatly increased the rigour of fingerprint property selection. However, despite the apparent rigour of the selection process, different source apportionments can be obtained using different combinations of fingerprint signatures (e.g. Laceby *et al.* 2015; Pulley *et al.* 2015b) and further refinement of the selection process is required.

Mixing models

The use of numerical mixing (or unmixing) models and related techniques to provide quantitative estimates of the relative contributions of the various potential sources to a sediment sample represented a key advance in sediment source fingerprinting. Because these models are generally overdetermined, optimization routines are usually employed to derive the estimates of the source contributions (e.g. Yu and Oldfield 1989; Collins *et al.* 1997; Krause *et al.* 2003). These are commonly based on minimization of the difference between the observed and predicted property values. It is important that the goodness of fit of the mixing model should be objectively tested to ensure that the result obtained is meaningful. The mixing models have been modified by some workers to include the use of weightings for the individual properties included in the model and the use of prior information to restrict the potential range of the optimized source contributions (Collins *et al.* 1998, 2010a). Collins *et al.* (2010b) have also recently demonstrated that alternative optimization procedures, including local and global genetic algorithm (GA) routines, may offer advantages. Partial least-squares regression models have also been used to estimate source contributions (Poulenard *et al.* 2009). The solutions provided by these different approaches can result in different source apportionments and there is a need for further refinement of modelling approaches and testing of the robustness of the outcomes (e.g. Haddachi *et al.* 2014; Laceby and Olley 2015; Sherriff *et al.* 2015).

Conservative behaviour, sediment enrichment and depletion

The fingerprinting technique necessarily assumes that the tracer properties behave conservatively within the fluvial system and also that the properties of sediment and source material samples can be directly compared (Owens and Xu 2011; Walling 2013; Wilkinson *et al.* 2015). It is often difficult to confirm conservative behaviour directly, but in many studies a simple range test, which confirms that the property values for the target sediment fall within the range of those for the potential sources, has been used to identify non-conservative behaviour. Enrichment and depletion effects associated with particle size composition and organic matter content, and resulting from selective mobilization and transport, can invalidate direct comparison of sediment and source material properties. Approaches employed to deal with this potential problem include use of the same particle size fraction of source material and target samples (e.g. Collins *et al.* 1998; Walling 2005; Douglas *et al.* 2010; Hatfield and Maher 2009) and the incorporation of correction factors into the mixing model (e.g. He and Owens 1995; Collins *et al.* 1998; Russell *et al.* 2001; Motha *et al.* 2003, 2004; Juracek and Ziegler 2009). The efficacy of such correction factors is, however, difficult to test and has been questioned in recent papers (e.g. Smith and Blake 2014; Pulley *et al.* 2015a).

Uncertainty in source apportionment

In common with many areas of hydrological modelling, increasing attention has been directed to the uncertainty of the results generated by sediment source fingerprinting investigations in recent years. As a result, Monte Carlo techniques have been incorporated into mixing model optimization routines to take account of the uncertainty associated with source characterization and to propagate this uncertainty through to the final source ascription results (e.g. Franks and Rowan 2000; Krause *et al.* 2003; Motha *et al.* 2004; Collins and Walling 2007; Collins *et al.* 2010a; Laceby and Olley 2015). Such models commonly produce a frequency distribution of estimates of the contribution of a given source, which reflects the uncertainty, rather than a single value. Bayesian approaches have also been used (Small *et al.* 2004; Douglas *et al.* 2007; Fox and Papanicolaou 2008; Palmer and Douglas 2008; Nosrati *et al.* 2014). Different modelling approaches can produce different outcomes and the research community has yet to establish a standardized modelling procedure.

Additional targets

Discrete samples of suspended sediment collected from a river have traditionally provided the 'target' in sediment source fingerprinting. However, such studies have demonstrated that sediment provenance can vary significantly both within and between events and problems can arise in defining the overall importance of individual sources to the longer term sediment output from a watershed. The development of time-integrating trap samplers that are able to collect a time-integrated sample of suspended sediment automatically (Phillips *et al.* 2000; Russell *et al.* 2000) has provided a valuable means of overcoming this problem. Other approaches have involved the use of samples of deposited sediment collected from riparian areas, the channel bed or the surface of floodplains, as a surrogate for the suspended sediment transported by a stream or river over a longer period. In addition, some studies have investigated the source of other specific types of fluvial sediment, including the fine sediment accumulating in salmonid spawning gravels (e.g. Walling *et al.* 2003a), fine bed sediment (e.g. Collins and Walling 2007) and lake and estuarine sediments (e.g. Foster and Walling 1994; Pittam *et al.* 2009).

Temporal change in sediment sources

Although early work on source fingerprinting focused on tracing the source of contemporary suspended sediment loads, the same basic approach has also been applied to dated sediment deposits (e.g. using ^{137}Cs and $^{210}Pb_{un}$) from lakes and reservoirs and river floodplains to reconstruct changes in sediment source through time (e.g. Foster *et al.* 1998; Owens *et al.* 1999; Walling *et al.* 2003c; Hatfield and Maher 2009; Pittam *et al.* 2009; Collins *et al.* 2010c; Foster *et al.* 2012; Pulley *et al.* 2015b). When assessing the efficacy of improved land management strategies in controlling sediment mobilization and delivery, repeated sediment source tracing exercises can potentially provide valuable information on changing sediment sources to complement information on the changing magnitude of the sediment fluxes (e.g. Merten *et al.* 2010). Use of the fingerprinting approach to identify longer term changes in source contributions through time, based on sediment deposits, introduces the need for further assumptions related to the absence of both temporal change in source properties and post-depositional diagenetic changes in the properties of the deposited sediment (Foster and Lees 2000; see also the following section).

Figure 9.5 provides an example of the results of sediment source fingerprinting investigations undertaken in two study areas in the United Kingdom (Walling *et al.* 2008). These results demonstrate significant contrasts both within and between the two study areas. Further details of this and other recent studies that have successfully employed sediment source fingerprinting techniques are provided in Table 9.4. To date, this approach has been used primarily as a research tool, rather than as an operational tool. Only limited progress has been made towards establishing standardized procedures that could be applied on a more routine basis and it is important that attention should be directed towards such standardization in future work.

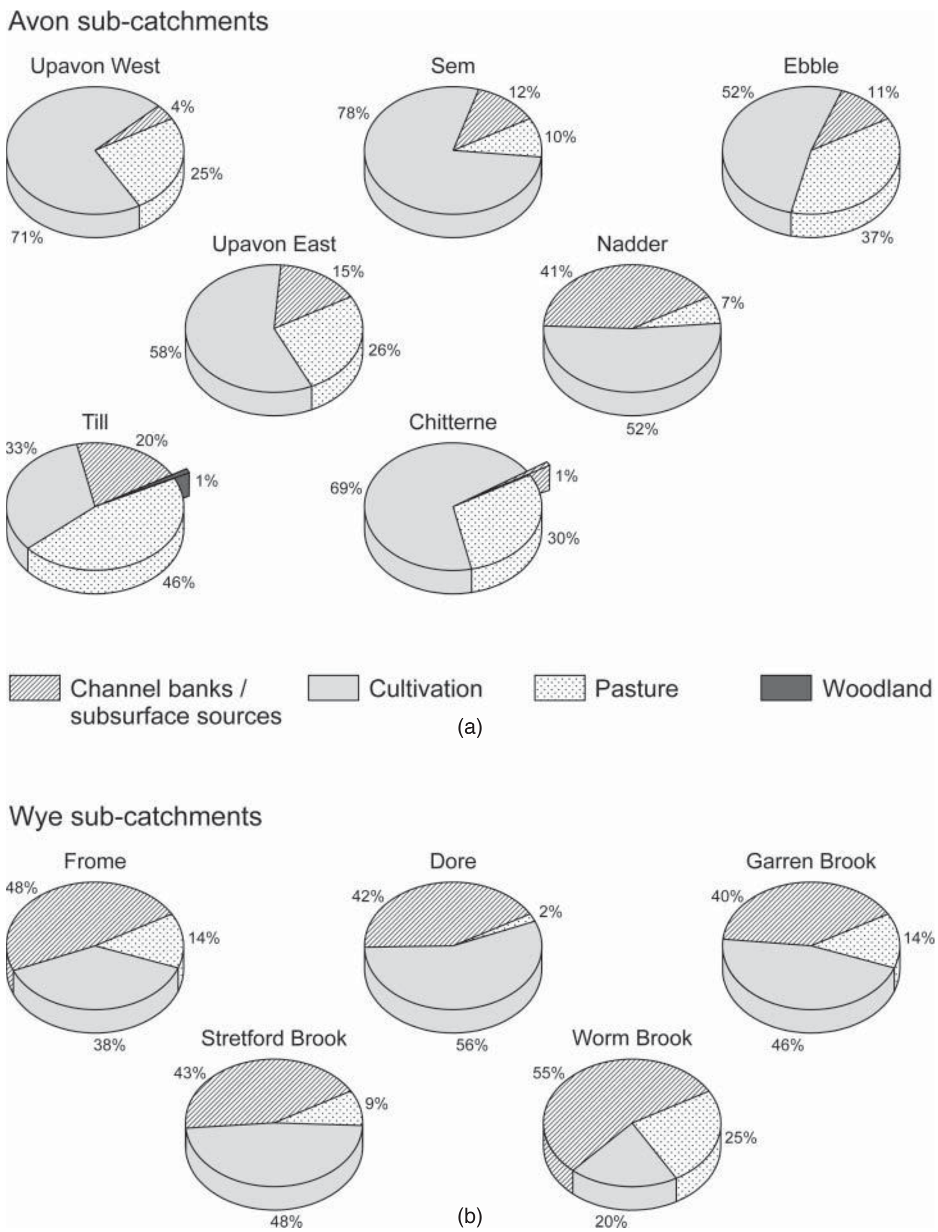


Figure 9.5 Mean source contributions to the annual sediment yields from selected sub-catchments of the Hampshire Avon, UK (a) and the Herefordshire Wye, UK (b), determined using source fingerprinting techniques. Adapted from Walling, 2008.

Table 9.4 Some examples of recent studies where source fingerprinting techniques have been successfully used to establish the primary suspended sediment sources within a catchment.

Reference	Location	Scale (km^2)	Sediment sources	Tracers used
Collins *et al.* (2010c)	Hampshire Avon sub-catchments	21–55	Channel/subsurface, agricultural topsoils, damaged road verges	ICP-MS, geochemistry
Douglas *et al.* (2008)	Fitzroy River basin, Queensland, Australia	144,000	Sub-areas defined by soil type/geology	Mineralogy, major and trace elements, isotope geochemistry
Devereux *et al.* (2010)	N.E. Anacostia River watershed, Maryland, USA	188	Hillslopes, stream channels and street residues	^{137}Cs, ^{40}K, total C and S, rare earth elements, trace elements
Fox and Papanicolaou (2007)	Upper Palouse basin, Idaho, USA	0.71	Hillslopes and Floodplains	δ^{13}C, δ^{15}N
Gellis *et al.* (2009)	Chesapeake Bay watershed, USA	109–156	Stream banks, construction sites, ditches, topsoil (agriculture, forests)	^{210}Pb,^{137}Cs, δ^{15}N, δ^{13}C, C, N and P
Hatfield and Maher (2009)	Bassenthwaite Lake catchment, UK	240	Topsoils and subsoils	Mineral magnetism
Martínez-Carreras *et al.* (2010a)	Attert catchment, Luxembourg	0.7–4.4	Topsoil (cultivated, forest, grassland), unmetalled roads and channel banks	Sediment colour, spectral reflectance
Mukundan *et al.* (2010)	North Fork Broad River, Georgia, USA	182	Stream channels, hillslopes, construction sites and unmetalled roads	^{137}Cs, δ^{15}N
Poulenard *et al.* (2009)	Albenche catchment, French Alps	9.9	Topsoils and channels	DRIFT, spectral reflectance
Walling *et al.* (2008)	Hampshire Avon and Herefordshire Wye catchments, UK	16–109	Channel banks and topsoil (cultivated, forest, pasture)	Radionuclides, metals, base cations and C, N and P concentrations

Reconstructing sediment accumulation rates, yields, sources and budgets

In a seminal paper, Oldfield (1977) argued that lake sediments preserve records of inputs from contributing catchments and can provide an uninterrupted record of change in catchment response over the lifespan of a lake or reservoir. Since the publication of this paper, there has been a major growth in the application of palaeoenvironmental reconstruction in geomorphological investigations. These applications include the reconstruction of sedimentation rates, sediment yields, sediment sources and sediment budgets to elucidate the relative significance of climate change, human activity and catchment connectivity, and reconstruction of the history of river flooding, earthquake-induced landslide activity, gullying, blanket peat erosion and the reworking of sediment by the rapid rise and fall of reservoir water levels (e.g. Curr 1995; Owens *et al.* 1997; Zolitschka 1998; Hyatt and Gilbert 2000; Nesje *et al.* 2001; Foster *et al.* 2002; Lamoureux 2002; Yeloff *et al.* 2005; Chiverrell *et al.* 2008; Couch and Eyles 2008; Koi *et al.* 2008; Hatfield and Maher 2009; Foster and Rowntree 2012; Rowntree and Foster 2012; Pulley *et al.* 2015b).

^{210}Pb chronologies derived from the methods outlined in the section 'Gamma-emitting radionuclides' allow changes in rates of sediment accumulation to be calculated for different depositional environments (colluvium, floodplains, lakes, reservoirs and estuaries). Information on the chronology of sediment accumulation in lakes and reservoirs can be combined with information on the volume and bulk density of the deposits, to estimate the mass of sediment deposited, and thus to reconstruct changes in the sediment yield from the contributing catchment. Such sediment yield reconstructions require a greater sampling effort, since a chronology derived from a single core is unlikely to be representative of the whole lake or reservoir basin (Dearing 1986; Dearing and Foster 1993; Foster 2006, 2010; Foster *et al.* 2011). Multiple cores could be dated, although this is usually prohibitively expensive, and the chronology derived from one or two dated cores is usually transferred to adjacent cores using one or more core correlation methods (e.g. matching downcore changes in mineral magnetic signatures, bulk density, particle size, organic matter content, pollen and diatoms). This allows the volume of sediment stored between isochrones to be calculated and conversion of these volumes to sediment mass is achieved using measured dry bulk density. Because much of the organic matter incorporated in the sediment deposits could be autogenic in origin, sediment yields are often reported on a minerogenic basis. Few lakes and reservoirs are 100% trap efficient and additional corrections for trap efficiency and changing trap efficiency through time are required (Dearing 1986; Foster 1995, 2010; Verstraeten and Poesen 2000). Examples of sediment yield reconstructions using ^{210}Pb and ^{137}Cs dating and multiple coring and core correlation for a UK reservoir and a South African farm dam are presented in Figure 9.6(a) and (b), respectively. The former shows short-lived increases in sediment yield associated with forest clearance and planting operations. The sediment yield is not corrected for organic matter content in Fig. 9.6(b), as loss on ignition is less than 5%. The range of sediment yield estimates (shown as lower, upper and average values in Fig. 9.6b) account for the range of calculated reservoir trap

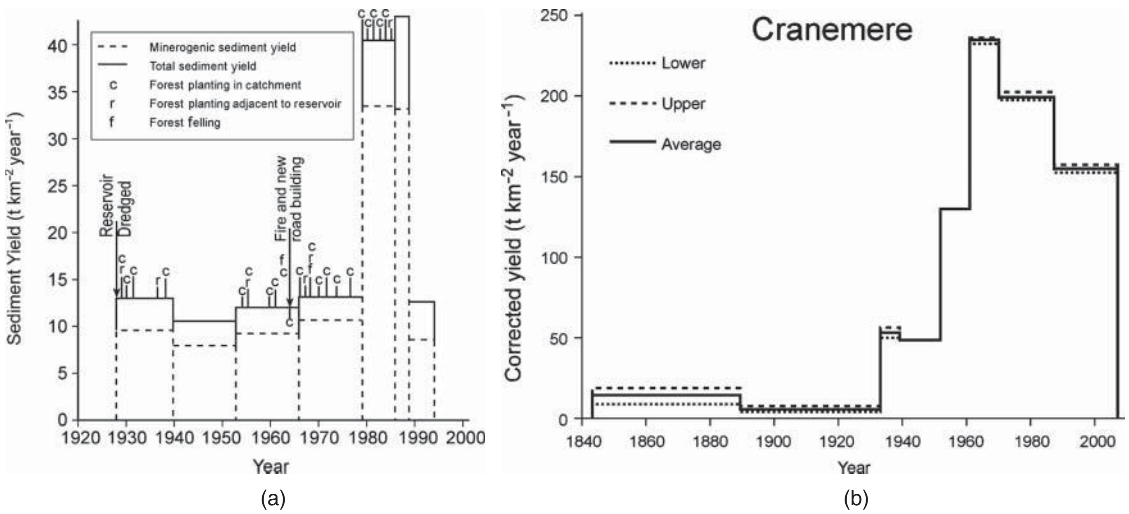


Figure 9.6 Sediment yield reconstructions for (a) Boltby reservoir, UK, showing total and minerogenic sediment yields and periods of catchment disturbance due to forest planting and felling (based on Foster and Lees 1999a) and (b) the Cranemere catchment on the plains of Camdeboo in the Eastern Cape' South Africa, showing total sediment yield (based on Foster and Rowntree 2012).

efficiencies over each period of sediment yield reconstruction. The dramatic increase in sediment yield in the early 1950s was probably caused by the construction of a causeway for a major road and the culverting of streams beneath the causeway, thereby reconnecting the upper catchment with the reservoir (Foster and Rowntree 2012).

In addition to the need to meet the various assumptions associated with the use of tracers outlined in the section 'Sediment source fingerprinting', further complications may arise when using them in a palaeoenvironmental context. Deposited sediment may retain its original geochemical, mineral magnetic and radionuclide signatures, although examples given in the section 'Sediment geochemistry' suggest that this is not the case for all tracers. Equally, when reconstructing sediment sources, it is not always possible to assume that the properties of the original source materials will have remained constant through time. In areas affected by significant atmospheric pollution, both geochemical and mineral magnetic signatures of catchment topsoils, for example, may have changed over the last century (e.g. Foster *et al.* 1990, 2002; Foster and Lees 1999b). Similar changes would be expected in the phosphorus concentration of agricultural topsoils. Short-lived radionuclides (e.g. ^{210}Pb, ^{137}Cs and ^{7}Be) are also unsuitable for long-term tracing as their activities will be influenced by radioactive decay. However, the long-lived gamma-emitting radionuclides identified in Table 9.1 will be suitable for tracing over periods spanning decades to centuries, provided that they are capable of discriminating potential catchment sources.

Early source tracing studies, involving sedimentary deposits, often used single tracers (e.g. Foster *et al.* 1990; Foster and Walling 1994) and, where reconnaissance studies are being undertaken, environmental magnetism alone will frequently provide evidence of significant changes in sediment source, without necessarily providing an exact apportionment. Figure 9.7(a) shows that a mineral magnetic signature (χ_{lf}) changes significantly through time in the sediments accumulating in a South African farm dam. Period 0–1 has high χ_{lf} values, similar to those of samples taken from a dolerite fan at the head of the eastern contributing catchment. Period 1–2 has much lower values, similar to those of badlands in the catchment, but the values recover to pre-existing levels during period 2–3. Period 3–4 again shows a decline in χ_{lf} that is sustained until the most recent period (4–5). Significant increases in sediment yield are associated with the connection of severely eroded badlands to the main channel feeding the reservoir as a result of channel avulsion (Rowntree and Foster 2012).

Other recently published studies have used a range of geochemical, mineral magnetic and radionuclide signatures, and a similar statistical framework to that identified in the section 'Sediment source fingerprinting', in order to discriminate likely sediment sources and quantify changes in their relative contribution through time (e.g. O'Malley *et al.* 1996; Foster and Lees 1999b; Foster *et al.* 2005, 2007; Turnbull *et al.* 2008; Mahapatra *et al.* 2011). In Fig. 9.7(b), sediment yield from a small lake catchment over the past ~80 years was reconstructed using multiple coring and a combination of ^{137}Cs and ^{210}Pb dating

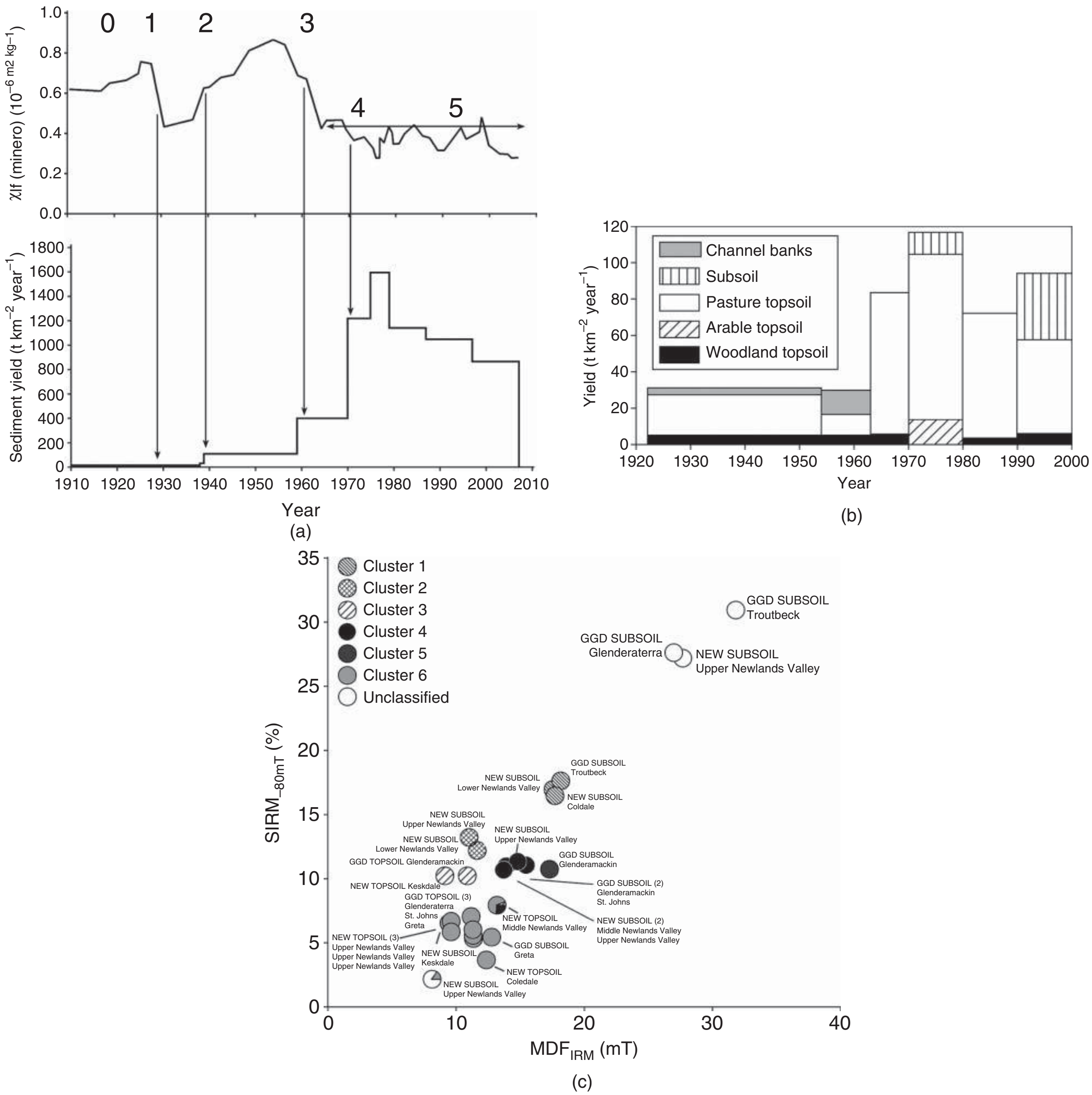


Figure 9.7 Reconstructing changes in sediment source based on lake and reservoir deposits. (a) Changing sediment sources and yields in the Ganora catchment, Eastern Cape, South Africa. Based on Foster and Rowntree (2012). (b) Changes in sediment sources associated with changing sediment yields in Kyre Pool, UK. Based on Foster (2006). (c) A bivariate MDF versus SIRM plot that discriminates six potential sediment source clusters contributing to Bassenthwaite Lake, UK. Adapted from Hatfield and Maher, 2009. See text for explanation.

(Foster *et al.* 2003). Source contributions over the 20th century to Kyre Pool were estimated using a linear unmixing model and a combination of long-lived radionuclide and mineral magnetic signatures (Foster 2006). The increase in sediment yield in the early 1960s, coupled with an increase in the amount of sediment derived from subsoil sources, was interpreted to reflect the installation of land drains in the catchment (Foster 2006).

Figure 9.7(c) demonstrates that magnetic signatures alone can provide a statistically robust method for source apportionment. The bivariate plot shows the result of using a fuzzy clustering approach based on an analysis of 15 soil profiles representing potential sources in the catchment. Six separate source groups are identified in Fig. 9.7(c) using this approach (Hatfield and Maher 2009).

9.4 Case study

Combining fallout radionuclide measurements and sediment source fingerprinting for sediment budgeting: Pang and Lambourn Catchments, United Kingdom

When used together and combined with measurements of the sediment yield at the catchment outlet provided by traditional monitoring techniques, fallout radionuclide measurements and sediment source fingerprinting can provide an effective basis for defining the main components of a catchment sediment budget (e.g. Walling *et al.* 2001, 2002b, 2006; Gellis and Walling 2011; Minella *et al.* 2014). Thus, for example, the measured sediment yield from a catchment can be combined with information on the source of the transported sediment provided by source fingerprinting, to apportion the sediment output according to its primary sources. Estimates of floodplain storage can be obtained using ^{137}Cs and/or $^{210}Pb_{un}$ measurements, and these can be added to the output flux to estimate the total sediment input to the channel system, which can again be apportioned to its primary sources. Fallout radionuclide measurements can be used to document gross and net rates of soil loss from the slope, and comparison of these estimates with estimates of sediment input to the channel system from slope sources provides a means of obtaining a first-order estimate of conveyance losses and storage associated with slope–channel transfer.

This approach, coupled with additional measurements of channel storage using the approach described by Lambert and Walling (1988), was used by Walling *et al.* (2006) to establish tentative sediment budgets for the Pang (166 km^2) and Lambourn (234 km^2) catchments in the United Kingdom. (See Chapter 16 for a review of sediment budgets generally.) These two catchments, located on the chalk of southern England, formed part of the Lowland Catchment Research Programme (LOCAR) funded by the UK Natural Environment Research Council (see http://catchments.nerc.ac.uk/). The location of the catchments on highly permeable strata with well-developed dry valley systems and the resulting dominance of groundwater flow mean that storm runoff is limited and that little sediment reaches the catchment outlets. However, there is evidence of relatively high rates of sediment mobilization and redistribution within the catchments and their sediment budgets are dominated by slope and slope to channel sediment sinks. These tentative sediment budgets are presented in Fig. 9.8.

The annual suspended sediment yields from these catchments were estimated to be very low at 3 and 4.6 t km^{-2} per year for the Pang and Lambourn catchments, respectively. In the absence of a well-developed floodplain and of significant overbank flows, overbank floodplain sedimentation could be treated as an insignificant component of the sediment budget. However, periodic measurements of channel storage of fine sediment at representative locations along the channel networks of the two catchments using the approached described by Lambert and Walling (1988) indicated that significant quantities of fine

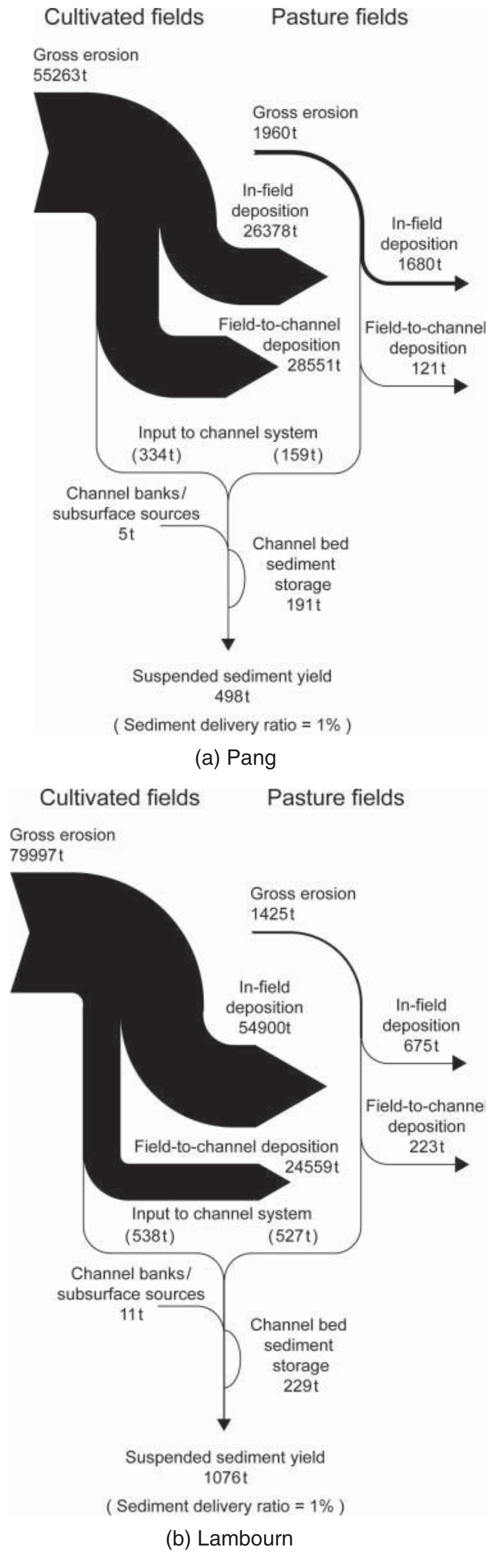


Figure 9.8 Catchment sediment budgets for (a) the Pang and (b) the Lambourn catchments in Berkshire, UK. The values indicated represent values of annual sediment flux and storage. Adapted from Walling, 2008.

sediment entered temporary storage in the river channels, but were flushed out by winter high flows. In view of its temporary nature, this storage did not represent a longer term or permanent sink, but it was estimated that of the order of 38% and 21% of the annual suspended sediment flux through the channel system was temporarily sequestered on the channel beds of the two catchments. Sediment source fingerprinting studies were undertaken in both catchments, with channel banks and surface soils from cultivated and pasture areas being identified as the primary potential sediment sources. These sources were sampled to characterize the source material properties and composite fingerprints capable of discriminating the three sources and incorporating geochemical, radiometric and organic properties were successfully identified. Time-integrating sediment samplers/traps based on the design of Phillips *et al.* (2000) were deployed at the catchment outlets to collect representative samples of the sediment output and these were used as the target for the sediment source apportionment. The source apportionments for the individual sampling intervals were combined to generate load-weighted mean contributions for the three potential sources. These data indicated that contributions from bank sources were low in both catchments (~1%) and that cultivated areas and pasture areas contributed 67% and 32%, respectively, of the sediment output from the Pang catchment and 50% and 49%, respectively, for the Lambourn catchment. When used in association with the measured sediment outputs, the source apportionment data provided a means of establishing the sediment inputs to the channels from the three primary sediment sources. Information on gross and net rates of sediment mobilization from the catchment slopes was assembled using ^{137}Cs measurements undertaken on cores collected from representative slope transects located in cultivated and pasture fields in both catchments. These data were extrapolated to the full catchment areas to provide estimates of gross and net erosion and thus information on sediment mobilization and storage within both cultivated and pasture fields on the catchment slopes and the onward transfer of sediment from these fields towards the channel network (Table 9.5). Comparison of the latter values with the estimate of sediment input to the channel systems of the two catchments, derived from the measured sediment yield and the sediment source fingerprinting, provided an estimate of the conveyance loss or storage associated with slope–channel transfer. The resulting sediment budgets shown in Fig. 9.8 clearly involve a number of uncertainties, particularly those associated with the estimates of slope–channel conveyance losses or storage, as these are based on the difference between the estimates of net erosion from the fields and sediment input to the channel system, rather than by direct measurement. These and other uncertainties associated with catchment sediment budgets are discussed by Kondolf and Matthews (1991). The budgets are, nevertheless seen as representing the key features of the sediment budgets of the chalk landscape of the study catchments, with its many dry valleys and low drainage density, which result in low slope–channel connectivity.

To consider the implications of these sediment budgets for the development of sediment management strategies within the Pang and Lambourn catchments, it is clear that measures to reduce the sediment output would need to target the slopes of the cultivated areas, since these represent the primary sediment source. A substantial reduction in sediment mobilization from the cultivated slopes would, nevertheless, be required to reduce sediment output from the catchments, since only a small proportion of the soil eroded from the cultivated area reaches the channel system. However, a small increase in the conveyance loss or deposition associated with field–channel transfer could result in an appreciable reduction in the sediment input to the channel system. A further reduction in the already low slope–channel connectivity should thus be seen as a priority target for control measures and improved management. Equally, the importance of in-field and field–channel storage in reducing the sediment input to the channels means that any change in the functioning of these sinks or stores, resulting in reduced deposition or perhaps remobilization of stored sediment, could potentially result in a major increase in the sediment outputs from the catchments in relative terms.

9.5 The prospect

The tools discussed in this chapter are often used by geomorphologists to date recent sedimentary deposits, document rates of soil and sediment redistribution, establish sediment sources, generate sediment budgets and reconstruct the functioning of past sediment systems. In focusing on common applications of radionuclide, mineral magnetic and geochemical signatures, particularly in relation to contemporary processes, we have

Table 9.5 Estimates of gross erosion, deposition and net erosion rates (t ha^{-1} year^{-1}) derived for pasture and cultivated fields in the Pang and Lambourn catchments.

Catchment	Land use	Gross erosion	Deposition	Net erosion	SDR* (%)
Pang	Cultivated	3.63	1.73	1.9	52
	Pasture	1.40	1.20	0.20	14
Lambourn	Cultivated	4.37	3.00	1.37	31
	Pasture	0.95	0.45	0.50	53

*Sediment delivery ratio, defined as the proportion of the sediment mobilized from the field by erosion that exits the field and is transported towards the channel network.

given only limited attention to the wider scene of reconstructing past landscape dynamics, where geomorphologists frequently collaborate with other scientists (especially palaeoecologists), in order to generate additional information about the environmental changes occurring within a catchment that may, for example, drive increases in soil erosion and sediment transport or additional information that could be used to validate chronologies derived from ^{210}Pb and ^{137}Cs dating. Multi-proxy reconstructions that add biological indicators (e.g. diatoms, pollen, chironomids, fungal spores) also have the potential to provide further evidence relating to changes in vegetation cover, grazing density and water quality in lakes or reservoirs and contributing rivers.

Although the use of fallout radionuclides for dating recent sediment deposits and documenting soil and sediment redistribution within the landscape is now fairly well established (e.g. Zapata 2002; IAEA 2014), the techniques involved in documenting soil and sediment redistribution are continuously being developed and improved (e.g. Walling *et al.* 2009; Walling 2010; Du and Walling 2012; Porto and Walling 2014; Porto *et al.* 2014) and further advances and refinements can be expected. In particular, the conjunctive use of $^{210}Pb_{un}$, ^{137}Cs and ^{7}Be to provide multi-temporal information on soil and sediment redistribution must be seen as offering considerable potential. The use of plutonium-239 and −240 ($^{239+240}$Pu) as an alternative to ^{137}Cs has also been shown to offer considerable potential (e.g. Everett *et al.* 2008; Alewell *et al.* 2014). These two fallout radionuclides are also a product of nuclear weapons testing in the 1950s and early 1960s, but they are characterized by much longer half-lives (24,110 and 6561 years, respectively). This means that they are not subject to the substantial decline in activity experienced by ^{137}Cs in recent years due to its short half-life, which in some locations in the southern hemisphere can be approaching the limit of detection. Furthermore, since they are derived from weapons testing, they afford a means of discriminating bomb- and Chernobyl-derived fallout. Sediment source fingerprinting techniques are, however, arguably at an earlier stage of development and require further synthesis and consolidation to produce general guidelines and standardized approaches. For example, there is currently a lack of guidance as to the optimum tracer properties to use in specific circumstances and the selection of an effective composite fingerprint commonly involves empirical testing of the discrimination afforded by a wide range of potential fingerprint properties. These techniques have benefited greatly from recent advances in automated geochemical analysis (e.g. mass spectrometers). These have greatly expanded the range of properties that can be analysed and which now include compound-specific stable isotopes and have greatly reduced the mass of soil sediment required for the analysis of many properties. Future advances can be expected to continue these trends. Leading in another direction, recent progress in the use of spectral reflectance and near-infrared (NIR) and mid-infrared (MIR) reflectance as surrogate measures of geochemical properties (e.g. Poulenard *et al.* 2009; Martínez-Carreras 2010a,b) offers considerable promise for the rapid and cheap characterization of soil and sediment properties and possibly even continuous in situ monitoring of suspended sediment properties.

The progressive development of these tools offers considerable potential for management applications, particularly for catchment management. For example, the ability to identify sources of fine sediment and changes in sediment sources through time can provide catchment managers with valuable support for developing targeted mitigation strategies and testing their effectiveness once implemented. Equally, reconstruction of the past sediment dynamics of a catchment can provide valuable information on historical reference conditions (Foster *et al.* 2011). There is a clear need to recognize the important role that fluvial geomorphology can play in understanding and valuing ecosystem services and underpinning the development of catchment management strategies (Gellis and Walling 2011).

References

Alewell, C., Meusburger, K., Juretzko, G., Mabit, L and Ketterer, M.E. 2014. Suitability of $^{239+240}$Pu and ^{137}Cs as tracers for soil erosion assessment in mountain grasslands. *Chemosphere* 103: 274–80.

Appleby, P.G. 2001. Chronostratigraphic techniques in recent sediments. In Last, W.M. and Smol, J.P., eds., *Tracking Environmental Change Using Lake Sediments. Volume 1: Basin Analysis, Coring and Chronological Techniques*, Dordrecht: Kluwer, pp. 171–204.

Appleby, P.G. 2008. Three decades of dating recent sediments by fallout radionuclides: a review. *Holocene* 18: 83–93.

Appleby, P.G. 2013. ^{210}Pb dating: thirty-five years on. *Journal of Paleolimnology* 49: 697–702.

Appleby, P.G. and Oldfield, F. 1978. The calculation of Pb-210 dates assuming a constant rate of supply of unsupported Pb-210 to the sediment. *Catena* 5: 1–8.

Appleby, P.G., Nolan, P., Gifford, D.W., Godfrey, M.J., Oldfield, F., Anderson, N.J. and Battarbee, R.W. 1986. Pb-210 dating by low background gamma counting. *Hydrobiologia* 141: 21–27.

Baskaran, M., ed. 2011. *Handbook of Environmental Isotope Geochemistry*, Heidelberg: Springer.

Benoit, G. and Hemond, H.F. 1991. Evidence for diffusive redistribution of ^{210}Pb in lake sediments. *Geochimica et Cosmochimica Acta* 55: 1963–l975.

Blake, W.H., Walling, D.E. and He, Q. 1999. Fallout beryllium-7 as a tracer in soil erosion investigations. *Applied Radiation and Isotopes* 51: 599–605.

Blake, W.H., Walling, D.E. and He, Q. 2002. Using cosmogenic beryllium-7 as a tracer in sediment budget investigations. *Geografiska Annaler* 84A: 89–102.

Blake, W.H., Wallbrink, P.H., Doerr, S.H., Shakesby, R.A. and Humphreys, G.S. 2006. Magnetic enhancement in wildfire-affected soil and its potential for sediment-source ascription. *Earth Surface Processes and Landforms* 31: 249–264.

Blake, W.H., Wallbrink, P.J., Wilkinson, S.N., Humphreys, G.S., Doerr, S.H., Shakesby, R.A. and Tomkins, K.M. 2009. Deriving hillslope sediment budgets in wildfire-affected forests using fallout radionuclide tracers. *Geomorphology* 104: 105–116.

Boyle, J.F., Dearing, J.A., Blundell, A. and Hannam, J.A. 2010. Testing competing hypotheses for soil magnetic susceptibility using a new chemical kinetic model. *Geology* 38: 1059–1062.

Brown, A.J. 1985. The potential use of pollen in the identification of suspended sediment sources. *Earth Surface Processes and Landforms* 10: 27–32.

Brenner, M., Schelske, C.L. and Kenney, W.F. 2004. Inputs of dissolved and particulate ^{226}Ra to lakes and implications for ^{210}Pb dating recent sediments. *Journal of Paleolimnology* 32: 53–66.

Caitcheon, G. 1993. Applying environmental magnetism to sediment tracing. In: *Tracers in Hydrology*, IAHS Publication 215, Wallingford: International Association of Hydrological Sciences, pp. 285–292.

Caitcheon, G. 1998. The significance of various sediment magnetic mineral fractions for tracing sediment sources in Killimicat Creek. *Catena* 32: 131–142.

Carter, J., Owens, P.N., Walling, D.E. and Leeks, G.J.L. 2003. Fingerprinting suspended sediment sources in a large urban river system. *Science of the Total Environment* 314–316: 513–534.

Chappell, A., Viscarro Rossel, R.A. and Loughran, R.A. 2011. Spatial uncertainty of ^{137}Cs-derived net (1950s–1990) soil redistribution for Australia. *Journal of Geophysical Research: Earth Surface* 116: F04015.

Chiverrell, R.C., Oldfield, F., Appleby, P.G., Barlow, D., Fisher, E., Thompson, R. and Wolff, G. 2008. Evidence for changes in Holocene sediment flux in Semer Water and Raydale, North Yorkshire, UK. *Geomorphology* 100: 70–82.

Collins, A.L. and Walling, D.E. 2004. Documenting catchment suspended sediment sources: problems, approaches and prospects. *Progress in Physical Geography* 28: 159–196.

Collins, A.L. and Walling, D.E. 2007. Sources of fine sediment recovered from the channel bed of lowland groundwater fed catchments in the UK. *Geomorphology* 88: 120–138.

Collins, A.L., Walling, D.E. and Leeks, G.J.L. 1997. Sediment sources in the Upper Severn catchment: a fingerprinting approach. *Hydrology and Earth System Sciences* 1: 509–521.

Collins, A. L., Walling, D.E. and Leeks, G.J.L. 1998. Use of composite fingerprints to determine the provenance of the contemporary suspended sediment load transported by rivers. *Earth Surface Processes and Landforms* 23: 31–52.

Collins, A L., Walling, D.E., Sichingabula, H.M. and Leeks, G.J.L. 2001. Suspended sediment source fingerprinting in a small tropical catchment and some management implications. *Applied Geography* 21: 387–412.

Collins, A.L., Walling, D.E., Webb, L. and King, P. 2010a. Apportioning catchment scale sediment sources using a modified composite fingerprinting technique incorporating property weightings and prior information. *Geoderma* 155: 249–261.

Collins, A.L., Zhang, Y., Walling, D.E., Grenfell, S.E. and Smith, P. 2010b Tracing sediment loss from eroding farm tracks using a geochemical fingerprinting procedure combining local and genetic algorithm optimisation, *Science of the Total Environment* 408: 5461–5471

Collins, A.L., Walling, D.E., Stroud, R.W., Robson, M. and Peet, L.M. 2010c Assessing damaged road verges as a suspended sediment source in the Hampshire Avon catchment, southern United Kingdom. *Hydrological Processes* 24: 1106–1122.

Couch, A.G. and Eyles, N. 2008. Sedimentary record of glacial Lake Mackenzie, Northwest Territories, Canada: implications for Arctic freshwater forcing. *Palaeogeography, Palaeoclimatology, Palaeoecology* 268: 26–38.

Crickmore, M.J., Tazioli, G.S., Appleby, P.G. and Oldfield F. 1990. *The Use of Nuclear Techniques in Sediment Transport and Sedimentation Problems*, IHP-III Project 5.2 SC-90/WS-49, Paris: UNESCO.

Curr, R. 1995. Magnitude and frequency of fluvial sediment transport determined from recent lake sediment cores. In: Foster, I.D.L., Gurnell, A.M. and Webb, B.W., eds., *Sediment and Water Quality in River Catchments*, Chichester: John Wiley & Sons, pp. 107–119.

Dearing, J.A. 1986. Core correlation and total sediment influx. In Berglund, B., ed., *Handbook of Holocene Palaeoecology and Palaeohydrology*, Chichester: John Wiley & Sons, pp. 247–270.

Dearing, J.A. 1999. Magnetic susceptibility. In: Walden, J., Oldfield, F. and Smith, J., eds., *Environmental Magnetism, a Practical Guide*, Technical Guide No. 6, London: Quaternary Research Association, pp. 35–62.

Dearing, J.A. 2000. Natural magnetic tracers in fluvial geomorpholgy. In: Foster, I.D.L., ed., *Tracers in Geomorphology*, Chichester: John Wiley & Sons, pp. 57–82.

Dearing, J.A. and Foster, I.D.L. 1993. Lake sediments and geomorphological processes: some thoughts. In: McManus, J. and Duck, R.W., eds., *Geomorphology and Sedimentology of Lakes and Reservoirs*, Chichester: John Wiley & Sons, pp. 5–14.

Dearing, J.A., Dann, R.J.L., Hay, K., Lees, J.A., Loveland, P.J., Maher, B.A. and O'Grady, K. 1996. Frequency-dependent susceptibility measurements of environmental materials. *Geophysics Journal International* 124: 228–240.

Devereux, O. H., Prestegaard, K.L., Needelman, B.A. and Gellis, A.C. 2010. Suspended-sediment sources in an urban watershed, Northeast Branch Anacostia River, Maryland. *Hydrological Processes*, 24: 1391–1403.

Douglas, G.B., Gray, C.M., Hart, B.T. and Beckett, R. 1995. A strontium isotopic investigation of the origin of suspended particulate matter (SPM) in the Murray–Darling river system, Australia. *Chemical Geology* 59: 3799–3815.

Douglas, G.B., Palmer, M. and Caitcheon, G. 2003. The provenance of sediments in Moreton Bay, Australia: a synthesis of major, trace element and Sr–Nd–Pb isotopic geochemistry, modelling and landscape analysis. *Hydrobiologia* 494: 145–152.

Douglas, G.B., Palmer, M., Caitcheon, G. and Orr, P. 2007. Identification of sediment sources to LakeWivenhoe, south-east Queensland, Australia. *Marine and Freshwater Research* 58: 793–810.

Douglas, G.B., Ford, P.W., Palmer, M.R., Noble, R.M., Packett, R.J. and Krull, E.S. 2008. Fitzroy River basin, Queensland, Australia. IV. Identification of flood sediment sources in the Fitzroy River. *Environmental Chemistry* 5: 243–257.

Douglas, G.B., Kuhnen, M., Radke, L.C., Hancock, G., Brooke, B., Palmer, M.J., Pietsch, T., Ford, P.W., Trefry, M.G. and Packett, R. 2010. Delineation of sediment sources to a coastal wetland in the Great Barrier Reef catchment: influence of climate variability and land clearing since European arrival. *Environmental Chemistry* 7: 190–206.

Du, P. and Walling, D.E. 2012. Using ^{210}Pb measurements to estimate sedimentation rates on river floodplains. *Journal of Environmental Radioactivity* 103: 59–75.

Everett, S.E., Tims, S.G., Hancock, G.J., Bartley, R. and Fifield, L.K. 2008. Comparison of Pu and ^{137}Cs as tracers of soil and sediment transport in a terrestrial environment. *Journal of Environmental Radioactivity* 99: 383–393.

Fairbanks, R.G., Mortlock, R.A., Chiu, T.-C., Cao, L., Kaplana, A., Guilderson, T.P., Fairbanks, T.W., Bloom, A.L., Grootes, P.M. and Nadeau, M.-J. 2005. Radiocarbon calibration curve spanning 0 to 50,000 years BP based on paired ^{230}Th/^{234}U/^{238}U and ^{14}C dates on pristine corals. *Quaternary Science Reviews* 24: 1781–1796.

Foster, I.D.L. 1991. *Environmental Pollution*, Oxford: Oxford University Press.

Foster, I.D.L. 1995. Lake and reservoir bottom-sediments as a source of soil erosion and sediment transport data in the UK. In: Foster, I.D.L., Gurnell, A.M. and B.W. Webb, eds., *Sediment and Water Quality in River Catchments*, Chichester: John Wiley & Sons, pp. 265–283.

Foster, I.D.L. 2006. Lakes in the sediment delivery system. In: Owens, P.N. and Collins, A.J., eds., *Soil Erosion and Sediment Redistribution in River Catchments*, Wallingford, CAB International, pp. 128–142.

Foster, I.D.L. 2010. Lakes and reservoirs in the sediment cascade. In: Burt, T.P. and Allison, R.J., eds., *Sediment Cascades: an Integrated Approach*, Chichester: John Wiley & Sons, pp. 345–376.

Foster, I.D.L. and Charlesworth, S.M. 1996. Heavy metals in the hydrological cycle: trends and explanation. *Hydrological Processes* 10: 227–261.

Foster, I.D.L. and Lees, J.A. 1999a. Changing headwater suspended sediment yields in the LOIS catchments over the last century: a palaeolimnological approach. *Hydrological Processes* 13: 1137–1153.

Foster, I.D.L. and Lees, J.A 1999b. Physical and geochemical properties of suspended sediments delivered to the headwaters of LOIS River Basins over the last 100 years: an analysis of lake and reservoir bottom-sediments. *Hydrological Processes* 13: 1067–1086.

Foster, I.D.L. and Lees, J.A. 2000. Tracers in geomorphology: theory and applications in tracing fine particulate sediments. In: Foster, I.D.L., ed., *Tracers in Geomorphology*, Chichester: John Wiley & Sons, pp. 3–20.

Foster, I.D.L. and Rowntree, K.M. 2012. Historic land management, rainfall and sediment yield changes in the semi-arid Karoo: a palaeoenvironmental reconstruction and interpretation of sediments accumulating in Cranemere Reservoir, Eastern Cape, South Africa. *Zeitschrift für Geomorphologie, Supplementband* 56(Suppl. 3): 131–146

Foster, I. D. L. and Walling, D.E. 1994. Using reservoir deposits to reconstruct changing sediment yields and sources in the catchment of the Old Mill Reservoir, South Devon, UK, over the past 50 years. *Hydrological Sciences Journal* 39: 347–368.

Foster, I.D.L., Dearing, J.A. Grew, R. and Orend, K. 1990. The sedimentary data base: an appraisal of lake and reservoir sediment based studies of sediment yield. In: Walling, D.E., Yair, A. and Berkowicz, S., eds., Erosion, Transport and Deposition Processes*, IAHS Publication 189*, Wallingford: International Association of Hydrological Sciences, pp. 19–43.

Foster, I.D.L., Lees, J.A., Owens, P.N. and Walling, D.E. 1998. Mineral magnetic characterizations of sediment sources from an analysis of lake and floodplain sediments in the catchments of the Old Mill reservoir and Slapton Ley, South Devon, UK. *Earth Surface Processes and Landforms* 23: 685–703.

Foster I.D.L., Lees, J.A., Jones, A.R., Chapman, A.S., Turner, S.E. and Hodgkinson, R.A. 2002. The possible role of agricultural land drains in sediment delivery to a small reservoir, Worcestershire UK; a multiparameter tracing study. In: Dyer, F.J., Thoms, M.C., and Olley, J.J., eds., *The Structure, Function and Management Implications of Fluvial Sedimentary Systems*, IAHS Publication 276, Wallingford: International Association of Hydrological Sciences, pp. 433–442.

Foster, I.D.L., Chapman, A.S., Hodgkinson, R.M., Jones, A.R., Lees, J.A., Turner, S.E. and Scott, M. 2003. Changing suspended sediment and particulate phosphorus loads and pathways in underdrained lowland agricultural catchments, Herefordshire and Worcestershire, UK. *Hydrobiologia*, 494: 119–126.

Foster, I.D.L., Boardman, J., Keay-Bright, J. and Meadows, M.E. 2005. Land degradation and sediment dynamics in the South African Karoo. In: Horowitz, A.J. and Walling D.E., eds., Sediment Budgets 2*, IAHS Publication 292*, Wallingford: International Association of Hydrological Sciences, 207–213.

Foster, I.D.L., Mighall, T.M., Proffitt, H., Walling, D.E. and Owens, P.N. 2006. Post-Depositional ^{137}Cs mobility in the sediments of three shallow coastal lagoons, SW England. *Journal of Paleolimnology* 35: 881–895.

Foster, I.D.L., Boardman, J. and Keay-Bright, J. 2007. The contribution of sediment tracing to an investigation of the environmental history of two small catchments in the uplands of the Karoo, South Africa. *Geomorphology* 90(1–2): 126–143.

Foster, I.D.L., Oldfield, F., Flower, R.J. and Keatings, K. 2008. Trends in mineral magnetic signatures in a long core from Lake Qarun, Middle Egypt. *Journal of Palaeolimnology* 40(3): 835–849.

Foster, I.D.L., Collins, A.L., Naden, P.S., Sear, D.A., Jones, J.I. and Zhang, Y. 2011. The potential for paleolimnology to determine historic sediment delivery to rivers. *Journal of Paleolimnology* 45(2): 287–306

Foster, I.D.L., Rowntree, K.M., Boardman, J. and Mighall, T.M. (2012) Changing sediment yield and sediment dynamics in the Karoo uplands, South Africa; post-European impacts. *Land Degradation and Development* 23: 508–522.

Fox, J.F. and Papanicolaou, A.N. 2007. The use of carbon and nitrogen isotopes to study watershed erosion processes. *Journal of the American Water Resources Association* 43: 1047–1064.

Fox, J.F. and Papanicolaou, A.N. 2008. An un-mixing model to study watershed erosion processes. *Advances in Water Resources* 31: 96–108.

Franks, S.W. and Rowan, J.S. 2000. Multi-parameter fingerprinting of sediment sources: uncertainty estimation and tracer selection. In: Bentley, L.R., Sykes, J.F., Gray, W.G., Brebbia, C.A. and Pinder, G.F., eds., *Computational Methods in Water Resources XIII*, Rotterdam: Balkema, pp. 1067–1074.

Gellis, A.C. and Munkundan, R. 2013. Watershed sediment source identification: tools, approaches and case studies. *Journal of Soils and Sediments* 13: 1655–1657.

Gellis, A.C. and Walling, D.E. 2011. Sediment source fingerprinting (tracing) and sediment budgets as tools in targeting river and watershed restoration programs. In: Simon, A., Bennett, S.J. and Castro, J.M., eds., Stream Restoration in Dynamic Fluvial Systems: Scientific Approaches, Analyses and Tools*, Geophysical Monograph Series 194*, Washington, DC: American Geophysical Union, pp. 263–291.

Gellis, A.C., Hupp, C.R., Pavich, M.J., Landwehr, J.M., Banks, W.S.L., Hubbard, B.E., Langland, M.J., Ritchie, J.C. and Reuter, J.M. 2009. *Sources, Transport and Storage of Sediment at Selected Sites in the Chesapeake Bay Watershed*, Scientific Investigations Report 2008-5186, Reston, VA: US Geological Survey.

Gibbs, M. 2008. Identifying source soils in contemporary estuarine sediments: a new compound-specific isotope method. *Estuaries and Coasts* 31: 344–359.

Gilmore, G. 2008. *Practical Gamma-ray Spectrometry*, 2nd edn. , Chichester: John Wiley & Sons.

Golosov, V. N. 2002. Special considerations for areas affected by Chernobyl fallout. In Zapata, F., ed., *Handbook for the Assessment of Soil Erosion and Sedimentation Using Environmental Radionuclides*, Dordrecht: Kluwer, pp. 165–183.

Graham, M.C., Vinogradoff, S.I., Chipchase, A.J., Dunn, S.M., Bacon, J.R. and Farmer, J.G. 2006. Using size fractionation and Pb isotopes to study Pb transport in the waters of an organic-rich upland catchment. *Environmental Science and Technology* 40: 1250–1256.

Grimshaw, D.L. and Lewin, J. 1980. Source identification for suspended sediments. *Journal of Hydrology* 47: 151–162.

Gruszowski, K.E., Foster, I.D.L., Lees, J.A. and Charlesworth, S.M. 2003. Sediment sources and transport pathways in a rural catchment, Herefordshire, UK. *Hydrological Processes* 17: 2665–2681.

Haddachi, A., Olley, J. and Laceby, J.P. 2014. Accuracy of mixing models in predicting sediment source contributions. *Science of the Total Environment* 497–498: 139–152.

Hassan, N.M., Rasmussen, P.E., Dabek-Zlotorzynska, E., Celo, V. and Chen, H. 2007. Analysis of environmental samples using microwave-assisted acid digestion and inductively coupled plasma mass spectrometry: maximizing total element recoveries. *Water Air and Soil Pollution* 178 323–334.

Hatfield, R.G. and Maher, B.A. 2009. Fingerprinting upland sediment sources: particle size-specific magnetic linkages between soils, lake sediments and suspended sediments. *Earth Surface Processes and Landforms* 34: 1359–1373.

He, Q. and Owens, P.N. 1995. Determination of suspended sediment provenance using caesium-137, unsupported lead-210 and radium-226: a numerical mixing model approach. In: Foster, I.D.L., Gurnell, A.M. and Webb, B.W., eds.,l *Sediment and Water Quality in River Catchments*, Chichester: John Wiley & Sons, pp. 207–227.

He, Q. and Walling D.E. 1996. Use of fallout Pb-210 measurements to investigate longer-term rates and patterns of overbank sediment deposition on the floodplains of lowland rivers. *Earth Surface Processes and Landforms* 21: 141–154.

He, Q., Walling, D.E. and Owens, P.N. 1996. Interpreting the ^{137}Cs profiles observed in several small lakes and reservoirs in southern England. *Chemical Geology* 129: 115–131.

Higgitt, S.R., Oldfield, F. and Appleby, P.G. 1991. Land use change and soil erosion in the pollen, magnetic and chemical record of the sediments of the last 2000 years in the Petit Lac d'Annecy, Eastern France. *The Holocene* 1: 14–28.

Hounslow, M.W. and Morton, A.C. 2004. Evaluation of sediment provenance using magnetic mineral inclusions in clastic silicates: comparison with heavy mineral analysis. *Sedimentary Geology* 171: 13–36.

Hyatt, J.A. and Gilbert, R. 2000. Lacustrine sedimentary record of human-induced gully erosion and land-use change at Providence Canyon, southwest Georgia, USA. *Journal of Paleolimnology* 23: 421–438.

IAEA. 2014. Guidelines for Using Fallout Radionuclides to Assess Erosion and Effectiveness of Soil Conservation Strategies. *IAEA-TECDOC 1741*, Vienna: International Atomic Energy Agency, 226 pp.

Joshi, S.R. and Mudroch, A. 1988. Direct determination of geochronologically useful radionuclides in sediments by low-energy photon analysis. *Nuclear Instruments and Methods in Physics Research* A263: 529–536.

Juracek, K.E. and Ziegler, A.C. 2009. Estimation of sediment sources using selected chemical tracers in the Perry lake basin, Kansas, USA. *International Journal of Sediment Research* 24: 108–125.

Kaushal, S. and Binford, M.W. 1999. Relationship between C:N ratios of lake sediments, organic matter sources and historical deforestation in Lake Pleasant, Massachusetts, USA. *Journal of Paleolimnology* 22: 439–442.

Kirchner, G. 2011. ^{210}Pb as a tool for establishing sediment chronologies: examples of potentials and limitations of conventional dating models. *Journal of Environmental Radioactivity* 102: 490–494.

Klages, M.G. and Hsieh, Y.P. 1975. Suspended solids carried by the Galatin River of southwestern Montana: II. *Using mineralogy for inferring sources. Journal of Environmental Quality* 4: 68–73.

Koi, T., Hotta, N., Ishigaki, I., Matuzaki, N.,Uchiyama, Y. and Suzuki, M. 2008. Prolonged impact of earthquake-induced landslides on sediment yield in a mountain watershed: the Tanzawa region, Japan. *Geomorphology* 101: 692–702.

Kondolf, G. M. and Matthews, W.V.G. 1991. Unmeasured residuals in sediment budgets: a cautionary note. *Water Resources Research* 27: 2483–2486.

Krause, A.K., Franks, S.W., Kalma, J.D., Loughran, R.J. and Rowan, J.S. 2003. Multi-parameter fingerprinting of sediment deposition in a small gullied catchment in SE Australia. *Catena* 53: 327–348.

Krom, M.D., Stanley, J.D., Cliff, R.A. and Woodward, J.C. 2002. Nile River sediment fluctuations over the past 7000 yr and their key role in sapropel development. *Geology* 30: 71–74.

Laceby, J.P. and Olley, J. 2015. An examination of geochemical modelling approaches to tracing sediment sources incorporating distribution mixing and elemental correlations. *Hydrological Processes* 29: 1669–1685.

Laceby, J.P., McMahon, J., Evrard, O. and Olley, J. 2015. A comparison of geological and statistical approaches to element selection for sediment fingerprinting. *Journal of Soils and Sediments* 15: 2117–2131.

Lambert, C.P. and Walling, D.E. 1988. Measurement of channel storage of suspended sediment in a gravel-bed river. *Catena* 15: 65–80.

Lamoureux, S. 2002. Temporal patterns of suspended sediment yield following moderate to extreme hydrological events recorded in varved lacustrine sediments. *Earth Surface Processes and Landforms* 27: 1107–1124.

Le Pape, P., Quantin, C., Morin, G., Jouvin, D., Kieffer, I., Proux, O., Ghanbaja, J. and Ayrault, S. 2014. Zinc speciation in the suspended particulate matter of an urban river (Orge, France): influence of seasonality and urbanization gradient. *Environmental Science and Techenololgy* 48: 11901–11909.

Li, S., Lobb, D.A., Tiessen, K.H.D. and McConkey, B.G. 2009. Selecting and applying cesium-137 conversion models to estimate soil erosion rates in cultivated fields. *Journal of Environmental Quality* 39: 204–219.

Longmore, M.E., O'Leary, B.M. and Rose, C.W. 1983. Cs-137 profiles in the sediments of a partial-meromictic lake on Great Sandy Island (Fraser Island), Queensland, Australia. *Hydrobiologia* 103: 21–27.

Loughran, R.J., Wallbrink, P.J., Walling, D.E. and Appleby, P.G. 2002. Sampling methods. In Zapata, F, ed., *Handbook for the Assessment of Soil Erosion and Sedimentation Using Environmental Radionuclides*, Dordrecht: Kluwer, pp. 41–57.

Mabit, L., Bernard, C. and Laverdière, M.R. 2007. Assessment of erosion in the Boyer River watershed (Canada) using a GIS oriented sampling strategy and ^{137}Cs measurements. *Catena* 71: 242–249.

Mabit, L., Benmansour, M. and Walling, D.E. 2008. Comparative advantages and limitations of the fallout radionuclides ^{137}Cs, $^{210}Pb_{ex}$ and ^{7}Be for assessing soil erosion and sedimentation. *Journal of Environmental Radioactivity* 99: 1799–1807.

Mabit, L., Benmansour, M., Abril, J.M., Walling, D.E., Meusburger, K., Iurian, A.R., Bernard, C., Tarjan, S., Owens, P.N., Blake, W.H. and Alewell, C. 2014. Fallout lead-210 as a soil and sediment tracer in catchment sediment budget investigations: a review. *Earth Science Reviews* 138: 335–351.

Mahapatra, D.M., Chanakya H.N. and Ramachandra, T.V. 2011. C:N ratio of sediments in a sewage fed urban lake. *International Journal of Geology* 5: 86–92.

Maher, B.A. 1988. Magnetic properties of some synthetic sub-micron magnetites. *Geophysical Journal of the Royal Astronomical Society* 94: 83–96.

Maher, B.A., Watkins, S.J., Brunskill, G., Alexander, J. and Fielding C. 2009. Sediment provenance in a tropical fluvial and marine context by magnetic fingerprinting. *Sedimentology* 56: 841–861.

Martínez-Carreras, N., Udelhoven, T., Krein, A., Gallart, F., Iffly, J.F., Ziebel, J., Hoffmann, L., Pfister, L. and Walling, D.E. 2010a.The use of sediment colour measured by diffuse reflectance spectrometry to determine sediment sources: application to the Attert River catchment (Luxembourg). *Journal of Hydrology* 382: 49–63.

Martínez-Carreras, N., Krein, A., Udelhoven, T., Gallart, F., Iffly, J.F., Hoffmann, L., Pfister, L. and Walling, D.E. 2010b. A rapid spectral-reflectance-based fingerprinting approach for documenting suspended sediment sources during storm runoff events. *Journal of Soils and Sediments* 10: 400–413.

Matisoff, G., Bonniwell, E.C. and Whiting, P.J. 2002. Soil erosion and sediment sources in an Ohio watershed using beryllium-7, cesium-137 and lead-210. *Journal of Environmental Quality* 31: 54–61.

Merten, G.H., Minella, J.P.G., Moro, M., Maier, C., Cassol, E.A., Walling, D.E., Clarke, R.T. and Reichert, J.M. 2010. The effects of soil conservation on sediment yield and sediment source dynamics in a catchment in southern Brazil. In: Banasik, K., Horowitz, A.J., Owens, P.N., Stone, M. and Walling, D.E., eds., *Sediment Dynamics for a Changing Future*, IAHS Publication 337, Wallingford: International Association of Hydrological Sciences, pp. 59–67.

Mighall, T.M., Martínez Cortizas, A., Silva Sánchez, N., Foster, I.D.L., Singh, S., Bateman, M. and Pickin, J. (2014) Identifying evidence for past mining and metallurgy from a record of metal contamination preserved in an ombrotrophic mire near Leadhills, SW Scotland, UK *The Holocene* 24: 1719–1730.

Minella, J.P., Walling, D.E. and Merten, G.H. 2014. Establishing a sediment budget for a small agricultural catchment in southern Brazil, to support the development of effective sediment management strategies. *Journal of Hydrology* 519: 2189–2201.

Motha, J.A., Wallbrink, P.J., Hairsine, P.B. and Grayson, R.B. 2003. Determining the sources of suspended sediment in a forested catchment in southeastern Australia. *Water Resources Research* 39: 1056–1069.

Motha, J.A., Wallbrink, P.J., Hairsine, P.B. and Grayson, R.B. 2004. Unsealed roads as suspended sediment sources in an agricultural catchment in south-eastern Australia. *Journal of Hydrology* 286: 1–18.

Mukundan, R., Radcliffe, D.E., Ritchie, J.C., Risse, L.M. and Mckinley, R. 2010. Sediment fingerprinting to determine the source of suspended sediment in a southern Piedmont stream. *Journal of Environmental Quality* 39: 1328–1337.

Murphy, B.L. and Morrison, R.D., eds., 2007. *Introduction to Environmental Forensics*, 2nd edn., Burlington, MA: Elsevier Academic Press.

Murray, A.S., Marten, R., Johnston, A. and Martin, P. 1987. Analysis for naturally occurring radionuclides at environmental concentration by gamma spectrometry. *Journal of Radioanalytical and Nuclear Chemistry: Articles*, 115: 263–288.

Nesje, A., Dahl, S.O., Matthews, J.A. and Berrisford, M.A. 2001. A ~ 4500-yr record of river floods obtained from a sediment core in Lake Atnsjøen, eastern Norway. *Journal of Paleolimnology* 25: 329–342.

Nosrati, K., Govers, G., Semmens, B.X. and Ward, E.J. 2014. A mixing model to incorporate uncertainty in sediment fingerprinting. *Geoderma* 217: 173–180.

Oldfield, F. 1977. Lakes and their drainage basins as units of sediment-based ecological study. *Progress in Physical Geography* 1: 460–504.

Oldfield, F. 1991. Sediment magnetism: soil erosion, bushfires or bacteria? *Geology* 19: 1155–1156.

Oldfield, F. 1994. Toward the discrimination of fine grained ferrimagnets by magnetic measurements in lake and near-shore marine sediments. *Journal of Geophysical Research* 99: 9045–9050.

Oldfield, F. 2007. Sources of fine-grained magnetic minerals in sediments: a problem revisited. *The Holocene* 17: 1265–1271.

Oldfield, F. and Crowther, J. 2007. Establishing fire incidence in temperate soils using magnetic measurements. *Palaeogeography, Palaeoclimatology, Palaeoecology* 249: 362–369.

Olley, J.M., Murray, A.S., Mackenzie, D.H. and Edwards, K. 1993. Identifying sediment sources in a gullied catchment using natural and anthropogenic radioactivity. *Water Resources Research* 29: 1037–1043.

O'Malley, V.P., Abrajano, T.A. Jr, and Hellou, J. 1996. Stable carbon isotopic apportionment of individual polycyclic aromatic hydrocarbons in St. John's Harbour, Newfoundland. *Environmental Science and Technology* 30: 634–639.

Owens, P.N. and Xu, Z. 2011. Recent advances and future directions in soils and sediments research. *Journal of Soils and Sediments* 11: 875–888.

Owens, P.N., Walling, D.E., He, Q., Shanahan, J. and Foster, I.D.L. (1997) The use of caesium-137 measurements to establish a sediment budget for the Start catchment, Devon,UK. *Hydrological Sciences Journal* 42: 405–423.

Owens, P.N., Walling, D.E. and Leeks G.J.L. 1999. Use of floodplain sediment cores to investigate recent historical changes in overbank sedimentation rates and sediment sources in the catchment of the River Ouse, Yorkshire, UK. *Catena* 36: 21–47.

Palmer, M.J.and Douglas, G.B. 2008. A Bayesian statistical model for end member analysis of sediment geochemistry, incorporating spatial dependences. *Applied Statistics* 57: 313–327.

Papanicolaou, A.N., Fox, J.F. and Marshall, J. 2003. Soil fingerprinting in the Palouse basin, USA, using stable carbon and nitrogen isotopes. *International Journal of Sediment Research* 18: 278–284.

Pennock D.J. and Appleby, P.G. 2002. Site selection and sampling design. In Zapata, F., ed., *Handbook for the Assessment of Soil Erosion and Sedimentation Using Environmental Radionuclides*, Dordrecht: Kluwer, pp. 15–40.

Phillips, J.M., Russell, M.A. and Walling, D.E. 2000. Time-integrated sampling of fluvial suspended sediment: a simple methodology for small catchments. *Hydrological Processes* 14: 2589–2602.

Pittam, N.J., Foster, I.D.L. and Mighall, T.M. 2009. An integrated lake-catchment approach for determining sediment source changes at Aqualate Mere, Central England. *Journal of Paleolimnology* 42: 215–232.

Porto, P. and Walling, D.E. 2014. Use of ^{7}Be measurements to estimate rates of soil loss from cultivated land: testing a new approach applicable to individual storm events occurring during an extended period. *Water Resources Research* 50: 8300–8313.

Porto, P., Walling, D.E. and Callegari, G. 2005. Investigating sediment sources within a small catchment in southern Italy. In: Walling, D.E. and Horowitz, A.J., eds., *Sediment Budgets I*, IAHS Publication No. 291, Wallingford: International Association of Hydrological Sciences, pp. 113–122.

Porto, P., Walling, D.E. and Callegari, G. 2010. Using ^{137}Cs measurements to establish catchment sediment budgets and explore scale effects. *Hydrological Processes* 25: 886–900.

Porto, P., Walling, D.E. and Callegari, G. 2013. Using ^{137}Cs and ^{210}Pb$_{ex}$ measurements to investigate the sediment budget of a small forested catchment in southern Italy. *Hydrological Processes* 27: 795–806.

Porto, P., Walling, D.E., Alewell, C., Callegari, G., Mabit, L., Mallimo, N., Meusburger, K. and Zehringer, M. 2014. Use of a ^{137}Cs re-sampling technique to investigate temporal changes in soil erosion and sediment mobilisation for a small forested catchment in southern Italy. *Journal of Environmental Radioactivity* 137: 137–148.

Poulenard, J., Perrette, Y., Fanget, B., Quetin, P., Trevisan, D. and Dorioz, J.M. 2009. Infrared spectroscopy tracing of sediment sources in a small rural watershed (French Alps). *Science of the Total Environment* 407: 2808–2819.

Pulley, S., Foster, I.D.L. and Antunes, P. 2015a. The uncertainties associated with sediment fingerprinting suspended and recently deposited fluvial sediment in the Nene river basin. *Geomorphology* 228: 303–319.

Pulley, S., Foster, I.D.L. and Antunes, P. 2015b The application of sediment fingerprinting to floodplain and lake sediment cores: assumptions and uncertainties evaluated through case studies in the Nene Basin, UK. *Journal of Soils and Sediments* 15: 2132–2154.

Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Bertrand, C. *et al.* 2009. Quaternary Isotope Laboratory, Radiocarbon Calibration Program, University of Washington. http://radiocarbon.pa.qub.ac.uk/calib/calib.html (accessed 11 October 2015).

Ritchie, J.C. and Ritchie, C.A. 2008. *Bibliography of Publications of ^{137}Cesium Studies Related to Erosion and Sediment Deposition*, USDA-ARS Hydrology and Remote Sensing Laboratory Occasional Paper HRSL-2008-02. http://www.ars.usda.gov/Main/docs.htm?docid=8865 (accessed 11 October 2015).

Ritchie, J.C., Finney, V.L., Oster, K.J. and Ritchie, C.A. 2004. Sediment deposition in the flood plain of Stemple Creek watershed, northern California. *Geomorphology* 61: 347–360.

Robbins, R.A. 1978. Geochemical and geophysical application of radioactive lead. In: Nriagu, J.O., ed., *The Biogeochemistry of Lead in the Environment*, Amsterdam: Elsevier, pp. 286–383.

Rose, N.L., Harlock, S., Appleby, P.G. and Battarbee, R.W. 1995. Dating of recent sediments in the United-Kingdom and Ireland using spheroidal carbonaceous particle (S.C.P.) concentration profiles. *The Holocene* 5: 328–335.

Rowntree, K.M. and Foster, I.D.L. 2012. A reconstruction of historical changes in sediment sources, sediment transfers and sediment yields for a small, semi-arid Karoo catchment, South Africa. *Zeitschrift für Geomorphologie, Supplementband* 56(Suppl. 1): 87–100.

Russell, M.A, Walling, D.E. and Hodgkinson, R.A. 2000. Appraisal of a simple sampling device for collecting time-integrated fluvial suspended sediment samples. In: Stone, M., ed., *The Role of Erosion and Sediment Transport in Nutrient and Contaminant Transfer*, IAHS Publication No. 263, Wallingford: International Association of Hydrological Sciences, pp. 119–127.

Russell, M.A., Walling, D.E. and Hodgkinson, R.A. 2001. Suspended sediment sources in two small lowland agricultural catchments in the UK. *Journal of Hydrology* 252: 1–24.

Schuller, P., Iroumé, A., Walling, D.E., Mancilla, H.B., Castillo, A and Trumper, R.E. 2006. Use of beryllium-7 to document soil redistribution following forest harvest operations. *Journal of Environmental Quality* 35: 1756–1763.

Schuller, P., Walling, D.E., Sepúlveda, A., Castillo, A. and Pino, I. 2007. Changes in soil erosion associated with the shift from conventional tillage to a no-tillage system, documented using ^{137}Cs measurements. *Soil and Tillage Research* 94: 183–192.

Sepulveda, A., Schuller, P., Walling, D.E. and Castillo, A. 2007. Use of ^{7}Be to document soil erosion associated with a short period of extreme rainfall. *Journal of Environmental Radioactivity* 99: 35–49.

Sherriff, S.C., Franks, S.W., Rowan, J.S., Fenton, O. and Ó'hUallacháin, D. 2015. Uncertainty-based assessment of tracer selection, tracer non-conservativeness and multiple solutions in sediment fingerprinting using synthetic and field data. *Journal of Soils and Sediments* 15: 2101–2116.

Slattery, M.C., Walden, J. and Burt, T.P. 2000. Fingerprinting suspended sediment sources using mineral magnetic measurements: a quantitative approach. In: Foster, I.D.L., ed., *Tracers in Geomorphology*, Chichester: John Wiley & Sons, pp. 309–322.

Small, I.F., Rowan, J.S., Franks, S.W., Wyatt, A. and Duck, R.W. 2004. Sediment fingerprinting using a Bayesian approach yields a robust tool for environmental forensic applications. In Croft, D.A. and Pye, K., eds., *Environmental Forensics*, Geological Society Special Publication, London: Geological Society, pp. 235–248.

Smith, H.G. and Blake, W.H. 2014. Sediment fingerprinting in agricultural catchments: a critical re-examination of source discrimination and data corrections. *Geomorphology* 204: 177–191.

Talbot, M.R., Williams, M.A.J. and Adamson, D.A. 2000. Strontium isotope evidence for late Pleistocene reestablishment of an integrated Nile drainage network. *Geology* 28: 343–346.

Terry, J.P., Garimella, S. and Kostaschuk, R.A. 2002. Rates of floodplain accretion in a tropical island river system impacted by cyclones and large floods. *Geomorphology* 42: 171–182.

Thapalia, A., Borrok, D.M., Van Metre, P.C. and Wilson, J. 2015. Zinc isotopic signatures in eight lake sediment cores from across the United States. *Environmental Science and Technology* 49: 132–140.

Thompson, R. and Oldfield, F. 1986. *Environmental Magnetism*, London: Allen and Unwin.

Turnbull, L., Brazier, R.E., Wainwright, J., Dixon, L. and Bol, R. 2008. Use of carbon isotope analysis to understand semi-arid erosion dynamics and long-term semi-arid land degradation. *Rapid Communications in Mass Spectrometry* 22: 1–6.

Vane, C.H., Chenery, S.R., Harrison, I., Kim, A.W., Moss-Hayes, V and Jones, D.G. 2011. Chemical signatures of the Anthropocene in the

Clyde estuary, UK: sediment-hosted Pb, $^{207/206}$Pb, total petroleum hydrocarbon, polyaromatic hydrocarbon and polychlorinated biphenyl pollution records *Philosophical Transactions of the Royal Society, A: Mathematical, Physical and Engineering Sciences* 369: 1085–1111.

Verosub, K.L. and Roberts, A.P. 1995. Environmental magnetism: past, present and future. *Journal of Geophysical Research* 100(B2): 2175–2192.

Verstraeten, G. and Poesen, J. 2000. Estimating trap efficiency of small reservoirs and ponds: methods and implications for the assessment of sediment yield. *Progress in Physical Geography* 24: 219–251.

Walden, J. 1999. Remanence measurements. In: Walden, J., Oldfield, F. and Smith, J., eds., *Environmental Magnetism, a Practical Guide*, Technical Guide No. 6, London: Quaternary Research Association, pp. 63–88.

Walden, J., Oldfield, F. and Smith, J., eds. 1999. *Environmental Magnetism, a Practical Guide*, Technical Guide No. 6, London: Quaternary Research Association.

Walker, M. 2005. *Quaternary Dating Methods*. Chichester: John Wiley & Sons.

Wall, G.J. and Wilding, L.P. 1976. Mineralogy and related parameters of fluvial suspended sediments in northwestern Ohio. *Journal of Environmental Quality* 5: 168–173.

Wallbrink, P.J., Murray, A.S. and Olley, J.M. 1998. Determining sources and transit times of suspended sediment in the Murrumbidgee River, New South Wales, Australia, using fallout ^{137}Cs and ^{210}Pb. *Water Resources Research* 34: 879–887.

Wallbrink, P.J., Murray, A.S. and Olley, J.M. 1999. Relating suspended sediment to its original soil depth using fallout radionuclides. *Soil Science Society of America Journal* 63: 369–378.

Wallbrink, P.J., Walling, D.E. and He, Q. 2002. Radionculide measurement using HPGe gamma spectrometry. In: Zapata, F., ed., *Handbook for the Assessment of Soil Erosion and Sedimentation Using Environmental Radionuclides*, Dordrecht: Kluwer, pp. 67–96.

Walling D.E. 2005. Tracing suspended sediment sources in catchments and river systems. *Science of the Total Environment* 344: 159–184.

Walling, D.E. 2010. Using fallout radionuclides to investigate erosion and sediment delivery: some recent advances. In: Banasik, K., Horowitz, A.J., Owens, P.N., Stone, M. and Walling, D.E., eds., *Sediment Dynamics for a Changing Future*, IAHS Publication 337, Wallingford: International Association of Hydrological Sciences, pp. 3–16.

Walling, D.E. 2012. The use of radiochemical measurements to investigate soil erosion and sedimentation. In: Meyers, R.A., ed., *Encyclopedia of Sustainability Science and Technology*, New York: Springer.

Walling, D.E. 2013. The evolution of sediment source fingerprinting investigations in fluvial systems. *Journal of Soils and Sediments*, 13: 1658–1675.

Walling, D.E. and Collins, A.L. 2008. The catchment sediment budget as a management tool. *Environmental Science and Policy* 11: 136–143.

Walling, D.E. and He, Q. 1992. Interpretation of cesium-137 profiles in lacustrine and other sediments: the role of catchment-derived inputs. *Hydrobiologia* 235: 219–230.

Walling, D.E. and He, Q. 1993. Use of cesium-137 as a tracer in the study of rates and patterns of floodplain sedimentation. In: Peters, N.E., Hoehn, E., Leibundgut, C., Tase, N. and Walling, D.E., eds., Tracers in Hydrology, *IAHS Publication 215*, Wallingford: International Association of Hydrological Sciences, pp. 319–328.

Walling, D.E. and He, Q. 1997. Use of fallout ^{137}Cs in investigations of overbank deposition on floodplains. *Catena* 29: 263–282.

Walling, D.E. and He, Q. 1999a. Improved models for estimating soil erosion rates from cesium-137 measurements. *Journal of Environmental Quality* 28: 1–622.

Walling, D.E. and He, Q. 1999b. Using fallout lead-210 measurements to estimate soil erosion on cultivated land. *Soil Science Society of America Journal* 63: 1404–1412.

Walling, D.E. and Quine, T.A. 1991. Use of ^{137}Cs measurements to investigate soil erosion on arable fields in the UK: potential applications and limitations. *Journal of Soil Science* 42: 147–165.

Walling, D.E. and Woodward, J.C. 1992. Use of radiometric fingerprints to derive information on suspended sediment sources. In: Bogen, J., Walling, D.E. and Day, T.J., eds., *Erosion and Sediment Transport Monitoring Programmes in River Basins*, IAHS Publication 210, Wallingford: International Association of Hydrological Sciences, pp. 53–64.

Walling, D.E. and Woodward, J.C. 1995. Tracing sources of suspended sediment in river basins: a case study of the River Culm, Devon, UK. *Marine and Freshwater Research* 46: 327–336.

Walling, D.E. and Zhang, Y. 2010. A national assessment of soil erosion based on caesium-137 measurements. *Advances in Geoecology* 41: 90–97.

Walling, D.E., Peart, M.R., Oldfield, F. and Thompson, R. 1979. Suspended sediment sources identified by magnetic measurements. *Nature* 281: 110–113.

Walling, D.E., Rowan, J.S. and Bradley, S.B. 1989. Sediment-associated transport and redistribution of Chernobyl fallout radionuclides. In: Hadley, R.F. and Ongley, E.D., eds., *Sediment and the Environment*, IAHS Publication 184, Wallingford: International Association of Hydrological Sciences, pp. 37–45.

Walling, D.E., He, Q. and Blake,W. H. 1999a. Use of ^{7}Be and ^{137}Cs measurements to document short- and medium-term rates of water-induced soil erosion on agricultural land. *Water Resources Research* 35: 3865–3874.

Walling, D.E., Owens P.N. Leeks, G.J.L. 1999b. Fingerprinting suspended sediment sources in the catchment of the River Ouse, Yorkshire, UK. *Hydrological Processes* 13: 955–975.

Walling, D.E., Collins, A.L., Sichingabula, H.W. and Leeks, G.J.L. 2001. Integrated assessment of catchment sediment budgets. *Land Degradation and Development* 12: 387–415.

Walling, D.E., He, Q. and Appleby, P.G. 2002a. Conversion models for use in soil erosion, soil redistribution and sedimentation investigations. In: Zapata, F., ed., *Handbook for the Assessment of Soil Erosion and Sedimentation Using Environmental Radionuclides*, Dordrecht: Kluwer, pp. 111–164.

Walling, D.E., Russell, M.A., Hodgkinson, R.A. and Zhang, Y. 2002b. Establishing sediment budgets for two small lowland agricultural catchments in the UK. *Catena* 47: 323–353.

Walling, D.E., Collins, A.L. and McMellin, G.K. 2003a. A reconnaissance survey of the source of interstitial fine sediment recovered from salmonid spawning gravels in England and Wales. *Hydrobiologia* 497: 91–108.

Walling, D.E., Collins, A.L. and Sichingabula, H.M. 2003b. Using unsupported lead-210 measurements to investigate soil erosion and sediment delivery in a small Zambian catchment. *Geomorphology* 52: 193–213.

Walling, D.E., Owens, P.N., Foster, I.D.L. and Lees, J.A. 2003c. Changes in the fine sediment dynamics of the Ouse and Tweed basins in the UK, over the last 100–150 years. *Hydrological Processes* 17: 3245–3269.

Walling, D.E., Collins, A.L., Jones, P.A., Leeks, G.J.L. and Old, G. 2006. Establishing finegrained sediment budgets for the Pang and Lambourn LOCAR catchments, UK. *Journal of Hydrology* 330: 126–141.

Walling, D.E., Collins, A.L. and Stroud, R.W. 2008. Tracing suspended sediment and particulate phosphorus sources in catchments. *Journal of Hydrology* 350: 274–289.

Walling, D.E., Schuller, P., Zhang, Y. and Iroumé, A. 2009. Extending the timescale for using beryllium 7 measurements to document soil redistribution by erosion. *Water Resources Research* 45: W02418.

Walling, D.E., Porto, P., Zhang, Y. and Du, P. 2014 Upscaling the use of fallout radionuclides in soil erosion and sediment budget investigations: addressing the challenges. *International Soil and Water Conservation Research* 2: 1–21.

Wiederhold, J.G. 2015. Metal stable isotope signatures as tracers in environmental geochemistry. *Environmental Science and Technology* 49: 2606–2624.

Wilkinson, S.N., Olley, J.M., Furuichi, T., Burton, J. and Kinsey-Henderson, A.E. 2015. Sediment source tracing with stratified sampling and weightings based on spatial gradients in soil erosion. *Journal of Soils and Sediments* 15: 2038–2051.

Wilson, C.G., Matisoff, G. and Whiting, P.J. 2003. Short-term erosion rates from a ^{7}Be inventory balance. *Earth Surface Processes and Landforms* 28: 967–977.

Wintle, A. and Murray, M. 2006. A review of quartz optically stimulated luminescence characteristics and their relevance in single-aliquot regeneration dating protocols. *Radiation Measurements* 41: 369–391.

Withers, P.J.A., Edwards, A.C. and Foy, R.H. 2001. Phosphorus cycling in UK agriculture and implications for phosphorus loss from soil. *Soil Use and Management* 17: 139–149.

Yeloff, D.E., Labadz, J.C., Hunt, C.O., Higgitt, D.L. and Foster, I.D.L. 2005. Blanket peat erosion and sediment yield in an upland reservoir catchment in the southern Pennines, UK. *Earth Surface Processes and Landforms* 30: 717–733.

Yu, L. and Oldfield, F. 1989. A multivariate mixing model for identifying sediment sources from magnetic measurements. *Quaternary Research* 32: 168–181.

Zaborska, A., Carroll, J., Papucci, C. and Pempkowiak, J. 2007. Intercomparison of alpha and gamma spectrometry techniques used in ^{210}Pb geochronology. *Journal of Environmental Radioactivity* 93: 38–50.

Zapata, F. 2002. *Handbook for the Assessment of Soil Erosion and Sedimentation Using Environmental Radionuclides*, Dordrecht: Kluwer.

Zapata, F., Garcia-Agudo, E., Ritchie, J.C. and Appleby, P.G. 2002. Introduction. In: Zapata, F., ed., *Handbook for the Assessment of Soil Erosion and Sedimentation Using Environmental Radionuclides*, Dordrecht: Kluwer, pp. 1–13.

Zhang, X. and Walling, D.E. 2005. Characterizing land surface erosion from cesium-137 profiles in lake and reservoir sediments. *Journal of Environmental Quality* 34: 514–523.

Zolitschka, B. 1998. A 14,000 year sediment yield record from Western Germany based on annually laminated lake sediments, *Geomorphology* 22: 1–17.

CHAPTER 10

Vegetation as a tool in the interpretation of fluvial geomorphic processes and landforms

Cliff R. Hupp[1], Simon Dufour[2] and Gudrun Bornette[3]

[1]US Geological Survey, Reston, VA, USA
[2]Université de Rennes 2, UMR CNRS LETG Rennes COSTEL, Rennes, France
[3]Université de Bourgogne Franche-Comté, Campus de la Bouloie, UMR CNRS 6249 Chronoenvironnement, Besançon, France

10.1 Introduction

The years since the first version of this chapter was published has seen an enormous increase in the number of publications that can be considered ecogeomorphic in nature (Hupp and Osterkamp 2013); most of these involve vegetation. Studies at the interface of ecology and geomorphology date back almost to the origin of the word 'ecology', first coined by Ernst Haeckel around 1870 (Odum 1971). Early ecologists clearly documented the value of using geomorphic form and process as a partial explanation of species distributions (e.g. Cowles 1901), which fundamentally defines the field, ecology. These early examples did not, however, lead to an immediate expansion in studies examining the interface of ecology and geomorphology. Since the end of the 19th century and prior to the very late 20th century, studies explicitly bridging the fields of ecology and geomorphology were, instead, sparse but included the important papers of Olson (1958) and Hack and Goodlett (1960).

Subsequent appreciation of interaction at this interface has led to the development of a field of endeavour termed 'biogeomorphology' (Viles 1988). Much more recently, a wide array of papers, books and symposia have attempted to integrate squarely the fast-emerging, now increasingly distinct field of biogeomorphology. Ecogeomorphology is essentially synonymous with biogeomorphology and perhaps is better reflective of the field (Hupp *et al.* 1995; Wheaton *et al.* 2011; Butler and Hupp 2013). Organisms from bacteria and their effects on weathering (Viles 1995) to woody plants and their effects on sedimentation and channel dynamics (Hupp and Osterkamp 1996; Osterkamp and Hupp 2010) may play a profound role in geomorphic processes. Early ecologists understood the need to document geomorphic form and process to explain plant species distributions (Cowles 1901). Thus, the relation between vegetation and geomorphic processes has now been acknowledged for more than a century. The field has expanded at such a rapid rate in the past 30 years that it would be impossible to treat fairly all phytogeomorphic studies with implications for vegetative tools within the confines of a single book chapter. Readers are directed to several symposia to observe the tremendous increase in the number of ecogeomorphic publications, including symposia proceedings (Thornes 1990; Hupp *et al.* 1995; Bennett and Simon 2004; Viles *et al.* 2008; Hession *et al.* 2010; Wheaton *et al.* 2011) and other papers where vegetation reflects geomorphic change and adjustment (Gurnell and Gregory 1995; Corenblit *et al.* 2008; Naylor 2005; Corenblit and Steiger 2009; Phillips 2009). The rise of this interest in ecogeomorphology and detailing of the causative impacts of geomorphic form and process on vegetation and vice versa have allowed the development of tools using vegetative responses to measure geomorphic processes quantitatively.

Vegetation can be used as a tool for geomorphic interpretation in several major ways, which will be exemplified after a general overview: (i) through dendrogeomorphic analyses (tree rings) to estimate the timing of important geomorphic events including floods and mass wasting and to estimate rates of erosion and sedimentation; (ii) through the documentation and interpretation of species distributional patterns that are established in response to prevailing hydrogeomorphic conditions; and (iii) through the role that it plays, depending on size, shape and growth form, in flow rates and subsequent erosion and deposition processes.

10.2 Scientific background: plant ecological–fluvial geomorphic relations

On the one hand, the community organization and dynamics of vegetation on river margins are strongly governed by fluvial geomorphic processes and landforms, which are largely created and maintained by fluctuations of water discharge. The likelihood of a given species vigorously growing on a particular landform is a function of the suitability of the site for germination and establishment and the ambient environmental conditions at the site that permit persistence at least until reproductive age (Grubb 1977; Zimmermann and Thom 1982; Hupp and Osterkamp 1996). On the other hand, aquatic and

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

riparian vegetation may affect fluvial geomorphic processes in various ways (Piégay and Gurnell 1997; Gregory *et al.* 2003; Corenblit *et al.* 2007, 2011): (i) by providing flow resistance, (ii) by increasing bank and other surface strength (e.g. roots), (iii) by facilitating channel-bar formation, (iv) by providing woody debris and (v) by facilitating deposition on bank-bench surfaces (Hickin 1984). These effects are generally related to the role that vegetation plays in sediment erosion and/or deposition. Thus, analysis of riparian vegetation ecology may lead to the use of indicative vegetation patterns as a tool in the interpretation and prediction of trends in fluvial geomorphic form and process.

Each floodplain or floodplain reach may be plotted on a curve that represents erosion and deposition processes (Fig. 10.1). The position along this curve depends on the dominant process (vertical axis) and on the size and number of disturbance patches (horizontal axis). For each river system or river reach, the amount of net aggradation versus erosion determines the position on the hypothetical curve simulating these two processes. Most temperate river floodplains (or floodplain reaches) can be plotted somewhere between these two polar situations where disturbance may be intense and the size of the deposition patches versus erosion patches is related to its location along the curve. A central equilibrium point corresponds to the situation where silted patches are as large as eroded patches and their numbers are also equal. This point reflects the highest habitat heterogeneity and potential biodiversity. In this case, a dynamic equilibrium (*sensu* Huston 1979; White and Pickett 1985) is reached, corresponding to a shifting mosaic of habitats and plant communities. Such situations are increasingly less common in temperate areas, owing to the frequent and heavy impacts of human activities that disrupt and drive fluvial processes towards one end or the other of the gradient. At the scale of an entire river, some reaches may be subjected to migrating aggradation but others to degradation (e.g. the impacts of channelization or dams).

Stream gradient (or power) also varies systematically along this conceptual gradient (Fig. 10.1). Where the stream gradient is high, erosional processes dominate (e.g. low-order high mountain cascades); conversely, along low-gradient reaches, depositional processes usually dominate (Hupp 2000). Thus, depending on the flow regime, equilibrium conditions and the highest biodiversity may shift along this conceptual gradient. Sediment sizes, stream gradient and channel pattern (meandering, cascading, straight) may change along the conceptual gradient to maintain adjusted conditions (Fig. 10.1).

Bornette *et al.* (2008) added a third dimension to this model to depict a gradient of patch turnover rate. This three-axis model incorporates the major physical limiting factors on riparian vegetation and provides a comprehensive explanation of plant life-history strategies. Readers are directed to Bornette *et al.* (2008) for a more in-depth discussion of disturbance frequency – patch turnover rates and their impacts on habitat dynamics and vegetation life-history strategies. Frequent relatively high-intensity disturbance should select species according to their tolerance for unstable environments

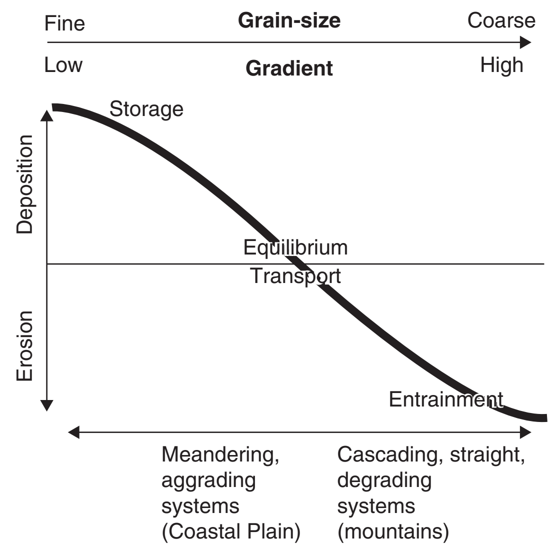


Figure 10.1 Pattern of sediment erosion versus deposition on river floodplains. Each river floodplain or river reach is characterized by deposition and/or erosion processes and the partitioning devoted to the two processes determines the position of the river reach along the curve. Note: depending on stream regime, equilibrium conditions may be on either side of the equal deposition/erosion point. Mountains and Coastal Plain are examples of a gradient (lowland to highland) from the Coastal Plain and Appalachian Mountains of southeastern United States.

(rapid turnover rates). Where disturbance is frequent and/or intense, physical interactions dominate and select for ruderal, disturbance-tolerant species (Bornette *et al.* 2008). As disturbance frequency decreases, biological interactions increase (competition, herbivory), which should increasingly select for competitive, stable site species. Also, as disturbance frequency lessens, other limiting factors, such as nutrient availability, fluxes of propagules and other biological interactions, become more important in determining riparian vegetation patterns (Bornette *et al.* 2008). Human alterations such as dam construction, land clearance with upland erosion and downstream aggradation, stream channelization or canal and levee construction may lead to channel incision or filling and large changes in sediment supply conditions depending on the geomorphic setting worldwide, and consequently may generate dramatic, usually rapid, regime shifts (Marston *et al.* 1995; Gurnell and Petts 2003; Rinaldi 2003; Gregory 2006; Hupp *et al.* 2009).

10.3 Vegetation as a tool: an overview

In spite of a voluminous literature demonstrating strong relationships between vegetation and hydrogeomorphic factors, publications that explicitly use vegetation as a tool are more limited. However, over the last few decades, vegetation has been successfully used to interpret various fluvial processes and patterns (Table 10.1). In bioclimatic and morphological contexts, vegetation can be used in various environments from boreal to tropical systems (e.g. Boucher *et al.* 2009; Pike and

Table 10.1 List of selected examples of vegetation usage as tools in fluvial geomorphology.

Discipline	Objectives (process or pattern characterized)	Type of measure/ principle	Biological support	Quantitative or qualitative	Scales (indicative)	Examples	Context
Hydromorphology	Reconstruction of past and changes in fluvial pattern and general conditions evolution (long term)	Species or genus identification	Pollen	Qual.	Spatial: reach Temporal: 100–10,000 years	Dobrowolski *et al.* 2010; Kasse *et al.* 2010; Nador *et al.* 2011; Turner *et al.* 2013 Brown 2002; Lespez *et al.* 2010 Gillson and Ekblom 2009 Willard *et al.* 2005 Rojo *et al.* 2012	Central Europe (Poland, Hungary, Germany) Western Europe (England, France) South Africa (Kruger Park) Eastern North America Andean piedmont, Argentina (arid and semi arid)
			Phytolith	Qual.	Spatial: reach Temporal: 100–10,000 years	Eichhorn *et al.* 2010; Garnier *et al.* 2013	West Africa (Sahelian, to Guineo–Congolian zones)
	Fluvial pattern classification	Vegetation: qualitative index of stabilization	Habitats/ community	Qual.	Spatial: reach Temporal: 1–300 years	Schumm 1985; Church 1992; Hupp 1992; Marston *et al.* 1995; Gurnell *et al.* 2001; Grams and Schmidt 2002; Ollero 2010; Belletti *et al.* 2015	Various contexts
	Qualified hydromorphological activity	Indicators of riparian dynamics (rate of vegetation establishment, diversity of habitats)	Habitats/ community	Qual./ Quant.	Spatial: reach Temporal: 1–300 years	Van Looy *et al.* 2008; Rollet *et al.* 2014	Large European western rivers
	Type of landform and dominant hydromorphological processes	Boundaries in vegetation types or species distribution indicate limit of a given landform and associated processes	Habitats/ community	Qual.	Spatial: landform Temporal: 1–500 years	Hupp and Osterkamp 1996 González Loyarte 2003	North American streams Piedmont of the central Andes (Argentina)
	Former channel connectivity	Community composition	Aquatic vegetation	Qual./ Quant.	Spatial: landform Temporal: 1–10 years	Amoros *et al.* 2000; Bornette and Arens 2002	Southeast France
	Specific stream power determination	Size of the lichen	Lichen*	Quant.	Spatial: reach Temporal: 10–300 years	Gob *et al.* 2003	Mediterranean rivers, South France
Geomorphology	Bank erosion intensity and rate	Dendrochronology, dendrogeomorphology (CWD and root exposure dating, inter- and intra-ring analysis)	Trees	Quant.	Spatial: landform to reach Temporal: 1–500 years	Malik 2006 Hitz *et al.* 2008 Stoffel *et al.* 2012 LaMarche 1968 Stotts *et al.* 2013; Dick *et al.* 2014	Southern Poland Switzerland (Alpine torrent) Patagonia Southwest USA Eastern USA

Sedimentation rate/depth	Dendrogeomorphology (tree ring)	Trees	Quant.	Spatial: landform Temporal: 1–500 years	Hupp and Bazemore 1993; Hupp and Simon 1991; Hupp *et al.* 1993; Magilligan and Stamp 1997	Southeast USA
					Friedman *et al.* 1996a	Great Plains USA
					Piégay *et al.* 2008; Provansal *et al.* 2010	Southeast France
Age of the surface	Dendrogeomorphology (tree ring)	Trees	Quant.	Spatial: landform Temporal: 1–500 years	Hupp 1988; Hupp *et al.* 1987	Southern and Pacific NW, USA
					Kochel *et al.* 1997	California, USA
					Fagot *et al.* 1989: Piégay *et al.* 2004	Southeast France (piedmont)
					Malik 2006	Southern Poland
					Nakamura *et al.* 2007	Japan
	Crown diameter	Trees	Quant.	Spatial: landform Temporal: 1–100 years	Beechie *et al.* 2006	Pacific NW, USA
	Size of the lichen	Lichen*	Quant.	Spatial: landform Temporal: 1–500 years	Harvey *et al.* 1984; Macklin *et al.* 1992	Upland rivers (England)
					Maas *et al.* 1998; Gob *et al.* 2003	Mediterranean rivers (Greece, southern France)
Age of cut-off	Dendrochronology (tree ring)	Trees	Quant.	Spatial: landform Temporal: 1–500 years	Harper 1912 Stella *et al.* 2011	Mississippi River, USA Sacramento River, USA
Date of disturbance events (floods, etc.)	Dendrochronology, dendrogeomorphology (inter- and intra-ring analysis)	Trees	Quant.	Spatial: landform Temporal: 1–500 years	Stoffel *et al.* 2008 Sigafoos 1964; Hupp 1988; Yanosky 1983	Switzerland (Alpine torrent) Eastern USA
Main channel mobility (degradation/aggradation)	Dendrochronology, dendrogeomorphology (ring width)	Trees	Qual./Quant.	Spatial: reach Temporal: 1–500 years	Dufour and Piégay 2008; Stella *et al.* 2013	Southeast France (piedmont)
					Hupp and Simon 1991; Hupp 1992	Southeast USA (channelized)
	Isotope analysis (from ring cellulose)			Spatial: reach Temporal: 1–500 years	Singer *et al.* 2012	Southeast France
	Community composition	Aquatic vegetation	Qual.	Spatial: landform Temporal: 1–10 years	Bornette *et al.* 1996	Southeast France

(*continued overleaf*)

Table 10.1 (*continued*)

Discipline	Objectives (process or pattern characterized)	Type of measure/ principle	Biological support	Quantitative or qualitative	Scales (indicative)	Examples	Context
Hydrology	Palaeofloods reconstruction	Size of the lichen	Lichen*	Quant.	Spatial: reach Temporal: 1–10,000 years	Maas *et al.* 2001	Andean mountain river, Bolivia
	Changes in hydrological regime	Dendrochronology (from scars on trees)	Trees	Quant./qual.	Spatial: reach Temporal: 1–500 years	Astrade and Begin 1997	Saone River, France (Oceanic)
	In-stream flow (mean discharge Jan–May)	Cottonwood branch growth	Trees	Quant.	Spatial: reach Temporal: 1–10 years	Willms *et al.* 1998	Rocky mountains, Alberta, Canada
	Frequency and magnitude of flood/high stage notably due to ice jam and spring floods	Dendrochronology (from scars on trees)	Trees	Quant.	Spatial: reach Temporal: 1–300 years	Uunila 1997; Boucher *et al.* 2009, 2011	Boreal rivers Canada
	Frequency of peak discharge	Presence and size of lichen	Lichen*	Quant.	Spatial: reach Temporal: 1–100 years	Gregory 1976	New England, Australia
	Estimation of flash flood discharge in an ungauged context	Dendrochronology (from scars on trees)	Trees	Quant.	Spatial: reach Temporal: 1–300 years	Ballesteros Cánovas *et al.* 2010	Mountain catchment, Spain (Mediterranean)
	General conditions (inundation duration, reference discharges such as bank full discharge)	Boundaries in vegetation types or species distribution indicate limit of a given flow frequency	Habitats/community	Qual./ quant.	Spatial: reach Temporal: 1–300 years	Osterkamp and Hupp 1984; Hupp and Osterkamp 1985	Eastern USA
						Wohl and Merritt 2008	Mountain streams in western USA, Nepal, New Zealand and Panama
						Pike and Scatena 2010	Subtropical mountains streams, northeastern Puerto Rico
						Follner and Henle 2006	River Elbe, Germany

*Lichen is composed of fungal and algal material and not vegetation, per se, but may be used similarly as a tool.

Scatena 2010) and from mountainous streams to lowland and coastal rivers (e.g. Stoffel *et al.* 2008; Wohl and Merritt 2008; Hupp *et al.* 2009). Beyond the diversity of approaches in terms of objectives and temporal scale (from 1 to 10,000 years), all are based on individuals (mainly trees), communities or pollen and applied at the landform or reach scale (Table 10.1).

Vegetation may be used as a geomorphic tool in three basic ways. First, vegetation may inform general fluvial geomorphic characters or processes acting at the reach scale (Schumm 1985; Church 1992), mainly in a qualitative or semiquantitative way (Table 10.1). For example, vegetation patterns on islands along braided reaches may distinguish between bar braided reaches and island braided reaches, thus indicating specific conditions in terms of hydrological regime, sediment supply, slope conditions and large woody debris (LWD) presence (Gurnell *et al.* 2001; Beechie *et al.* 2006; Belletti *et al.* 2015). The presence of pioneer habitats can also be used as an indicator of the global dynamic of the system and considered a good indicator of system integrity (Van Looy *et al.* 2008; Rollet *et al.* 2014). Such information can be integrated over time (decades) to reconstruct quantitatively fluvial landscape trajectory and hydrological and morphological changes by using images and old maps (Marston *et al.* 1995; Kondolf *et al.* 2007; Dufour *et al.* 2012). Vegetation patterns may also qualitatively date changes in fluvial systems over centuries and millennia by using pollen or phytolith (microscopic silica structures produced in the tissues of many plants) (Brown 2002; Eichhorn *et al.* 2010; Willard *et al.* 2005, 2011; Garnier *et al.* 2013). The distributional pattern of many plant species may be limited by their tolerance for specific types of disturbance and stress and consequently by tolerance for biotic interactions that prevail at this disturbance or stress level (Bornette *et al.* 2008). Riparian vegetation patterns (Fig. 10.2), even along highly altered streams, are indicative of present and ongoing fluvial forms and processes, while simultaneously reflecting stages of channel dynamics, for example, following incision, channel widening and/or narrowing (Hupp 1992; Marston *et al.* 1995; Friedman and Lee 2002; Hupp and Rinaldi 2007; Osterkamp *et al.* 2012).

Second, vegetation may also quantify more specific geomorphic processes such as sedimentation/erosion rates and/or landform age. For example, floodplain sedimentation/erosion regimes and rates can be quantified through dendrogeomorphological analysis (Sigafoos 1964; Hupp and Morris 1990; Magilligan and Stamp 1997; Ross *et al.* 2004; Piégay *et al.* 2008; Kroes and Hupp 2010). Main channel mobility can also be monitored using vegetation; for example, in the lateral dimension, Malik (2006) and Stoffel *et al.* (2012) studied bank erosion rate and Nanson and Beach (1977) characterized meander progradation. In the vertical dimension, degradation or aggradation of the main channel can be studied through riparian vegetation response. Indeed, in alluvial reaches, water table position is generally linked to main channel position and tree rings record changes in water availability. Thus, the analysis of the ring width (Dufour and Piégay 2008; Stella *et al.* 2013) and also of

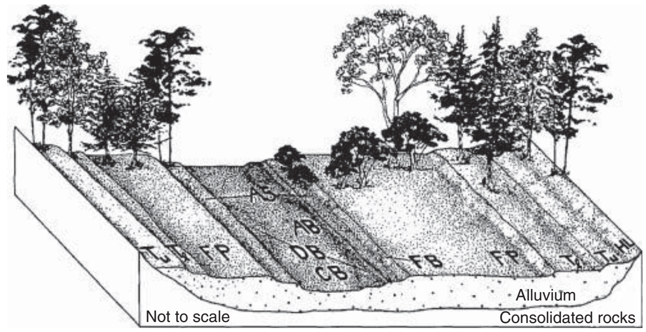


Figure 10.2 Generalized common fluvial geomorphic features along perennial streams (northern Virginia); from the lowest, the surfaces are CB, channel bed; DB, depositional bar; AB, active channel bank; AS, active channel shelf; FB, floodplain bank; FP, floodplain; Tl, lower terrace; Tu, upper terrace; Hl, hillslope. Vegetation species may characterize each of the fluvial surfaces. Source: Osterkamp and Hupp, 1984.

the oxygen isotopic signature in ring cellulose (Singer *et al.* 2012) provides a retrospective access to vertical main-channel mobility. At landform scale, some key parameters such as age can be accurately assessed through dendrochronology (Sigafoos and Hendricks 1961, Hupp *et al.* 1987; Pierson 2007; Stella *et al.* 2011), tree size (Beechie *et al.* 2006) or lichen dimension (Gob *et al.* 2003.) For example, a study by Beechie *et al.* (2006) showed that tree crown diameters were related to stand age (Fig. 10.3) and thus age of patch or disturbance.

Third, vegetation helps to quantify or characterize hydrological processes or parameters very useful for geomorphic studies. Vegetation reflects general flow regime because it may record mean hydrological conditions. This can be used to monitor change in terms of flow regime (Willms *et al.* 1998; Astrade 2005; Follner and Henle 2006) and also to evaluate the inundation duration at landform scale (Yanosky 1983; Osterkamp and Hupp 1984) and also specific events such as floods (Hupp 1988). Indeed, both flood frequency (Hupp 1987; Uunila 1997) and flood discharge (Ballesteros Cánovas *et al.* 2010) have been

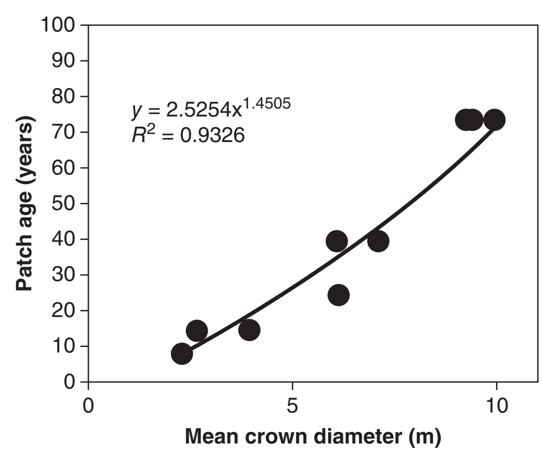


Figure 10.3 Relation between patch age and crown diameter. Maximum age was set to 100 years due to floodplain logging. Beechie *et al*, 2006. Reproduced with permission of Elsevier.

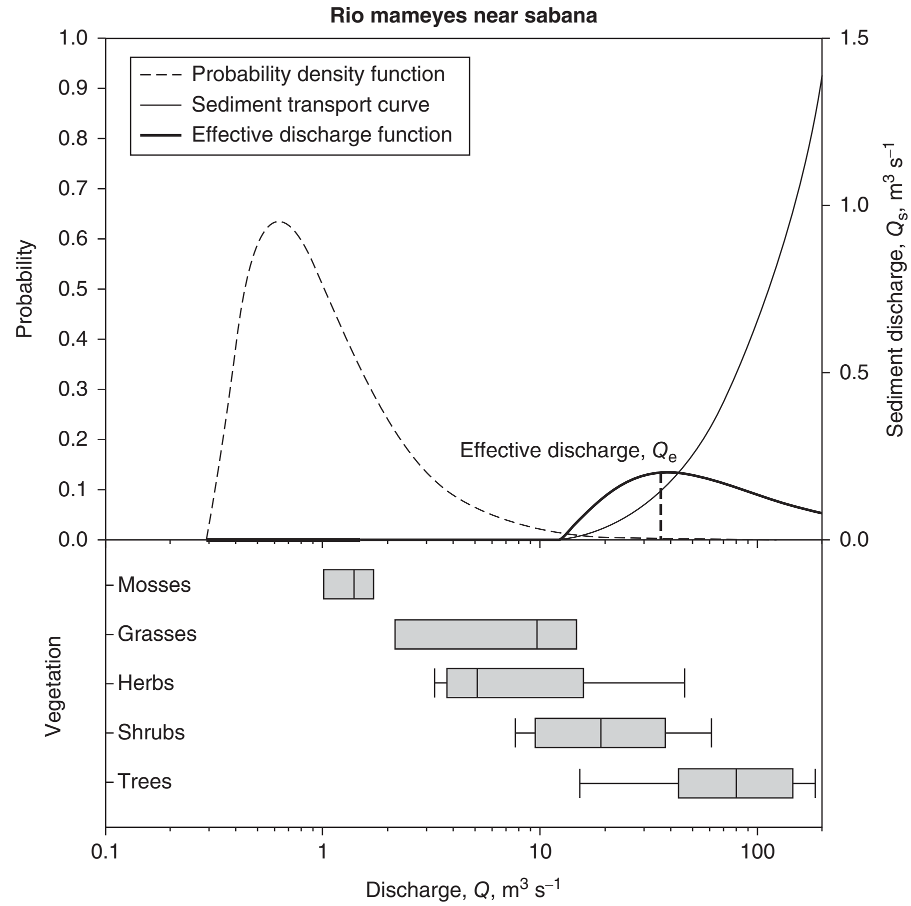


Figure 10.4 Relations among amount of discharge, discharge probability and effective discharge. Woody vegetation (shrubs and trees) occur on fluvial surfaces that are coincident with effective discharges, while herbaceous plants occur at lower discharges. Source: Pike and Scatena, 2010. Reproduced with permission of Elsevier.

measured using vegetation indicators. Sediment-transporting discharges, or effective discharge, have been tightly linked to riparian vegetation types (Pike and Scatena 2010) occurring at various elevations (stage of flow); they show that woody, rather than herbaceous, plants grow on surfaces that are correlated with the effective discharge (Fig. 10.4). Additionally, parameters such as stream power can also be evaluated using vegetation (Bendix and Hupp 2000) and lichen (Gob *et al.* 2003) measurement.

Among all these tools, dendrogeomorphology has a very important place in terms of frequency and variety of uses (see Table 10.1). Indeed, it provides quantitative information in a wide range of fluvial applications, including floods (Sigafoos 1964; Yanosky 1982; Hupp 1988; St. George and Nielsen 2003), floodplain deposition and aggradation (Sigafoos 1964; Hupp 1988; Hupp and Bazemore 1993; Strunk 1997; Friedman *et al.* 2005; Kroes and Hupp 2010; Merigliano *et al.* 2013), channel dynamics and mobility (Hupp 1992; Friedman *et al.* 1996a; Scott *et al.*1997; Dufour and Piégay 2008; Stella *et al.* 2013), estuary and lake shoreline dynamics (Begin and Filion 1995; Begin *et al.* 1991), saltwater intrusion along streams (Yanosky *et al.* 1995) and mountain glacier activity and debris flows (Sigafoos and Hendricks 1961; Hupp *et al.* 1987; Strunk 1997; Stoffel *et al.* 2005; Bollschweiler *et al.* 2007; Koch 2009; Bollschweiler and Stoffel 2010).

10.4 Dendrogeomorphology in fluvial systems

The use of tree-ring dating for the interpretation of geomorphic processes has become an increasingly common technique, so much so that an exhaustive list of dendrogeomorphic studies is well beyond the scope of this chapter. Several early classic papers that underscored the use of tree-ring information include Sigafoos (1964), Everitt (1968), Alestalo (1971), Helley and LaMarche (1973) and Schweingruber (1988). The term dendrogeomorphology was initially used by Shroder (1978) in his work on mass movement. Fairly extensive reviews can be found in several chapters of Jacoby and Hornbeck (1987). Two more recent reviews, one focusing on debris flows, rock falls and other hillslope processes by Stoffel *et al.* (2008) and the other with a focus on stream flow and channel geometry by Merigliano *et al.* (2013), provide a broad background of the topic. Many forms of natural hazards may be investigated using dendrogeomorphic approaches (Stoffel and Bollschweiler 2008).

Where historic records are short or lacking, tree-ring study may be the most accurate method for obtaining magnitude and frequency data over the past few hundred years. Botanical evidence, as a tool, in combination with geomorphic evidence allows for the interpretation and determination of the relative importance of various geomorphic processes. Beyond dating of specific episodic events, dendrogeomorphic analyses (Shroder 1978) permits the estimation of rates and amounts of important processes, including sediment erosion and deposition and channel mobility, which are the core of fluvial geomorphic work (Fig. 10.5), as exemplified by Sigafoos (1964). Note that dendrogeomorphology differs from dendrochronology such that once the tree ring(s) are identified as evidence of geomorphic activity; the rings towards the outside of the specimen are simply counted to determine age. Standard dendrochronological techniques that incorporate ring-width measurement to interpret geomorphic processes may also be useful, especially

Figure 10.5 Floodplain tree (Potomac River, Virginia) showing evidence of three geomorphically significant floods. First, a flood partially pushed over an ash (Fraxinus pennsylvanica) tree (b, c) causing tilt sprout formation (a) and eccentric growth (b). A second flood deposited sediment to a point above (b) but below (a), causing a change from stem wood to root wood formation (c) and adventitious root formation (e). A third flood eroded sediment to present ground surface (d) and caused a change from root wood to stem wood formation (f). See Section 10.3 for definition of types of botanical evidence shown (a, b, c, etc.). Photograph by R.S. Sigafoos.

in documenting long-term hydrological trends (Phipps 1983; Astrade and Bégin 1997; Cleaveland 2000; Boucher *et al.* 2011; Singer *et al.* 2012).

Floods and inundation

Floods, from prolonged inundation characteristic of relatively large, low-gradient basins to high-gradient and short-period (flashy) destructive events, are, perhaps, the most important extrinsic factor in bottomland systems. Thus, the knowledge of flooding characteristics and magnitude/frequency information is of great utility to students of fluvial geomorphology. Dates of past floods may be determined from ages of trees on fluvial landforms, from scars and sprouts on flood-damaged stems (Sigafoos 1964; Hupp 1988), from differences in properties of wood anatomy related to flooding (Yanosky 1982; St. George and Nielsen 2003) and from root anatomy following erosive floods (Gärtner 2007; Stoffel *et al.* 2012).

Four basic types of botanical evidence of geomorphic events, floods in this case, are routinely used (Sigafoos 1964; Hupp 1988): (i) corrasion scars, (ii) adventitious sprouts, (iii) ring anomalies and (iv) tree age (Fig. 10.6). A description of each follows. It is assumed that all samples are cross-dated, which reduces problems associated with false or missing rings (Cleaveland 1980; Cleaveland and Stahle 1989). Many other natural and human disturbances can cause forms of botanical evidence; careful examination of specimens and replication are required to limit spurious interpretations.

Corrasion scars. Scars may be the most conspicuous evidence of past flooding on riparian trees and shrubs (Fig. 10.6a). Currently, the most reliable and accurate method of tree-ring-determined dating of floods is the analysis of increment cores or cross-sections through scars. These samples may yield the exact date (year) of an event and, because of differences in wood produced during the growing season, the season of the event also may be inferred. Corrasion destroys the cambium (wood-producing tissue) in the area of impact, thus growth is stopped in the damaged area and the event is recorded as undamaged tissue grows in annual increments around and over the scar. Maximum scar heights may also allow for the estimation of the peak stage of a flood or elevation of a debris flow. The dendrogeomorphic use of scars or wounding including resin duct formation is discussed at length with particular focus on coniferous species in Stoffel and Bollschweiler (2008).

Adventitious sprouts. Sprouts from inclined or broken stems are easily determined in the field and can date an event by coring at the base of the sprout. This type of evidence has the appearance of vertically growing sprouts from a tilted main stem (Figs 10.5 and 10.6b) that is usually trained in the downstream direction or has two or more sprouts growing from a split base (Fig. 10.6c). Flood training of riparian trees is a highly visible feature along streams subjected to periodic high-velocity floods. Use of the term sprout does not imply youth, as some sprouts may be fairly old. The age of sprouts is usually within 1 year of the age of the event; obviously, the first year (centre ring)

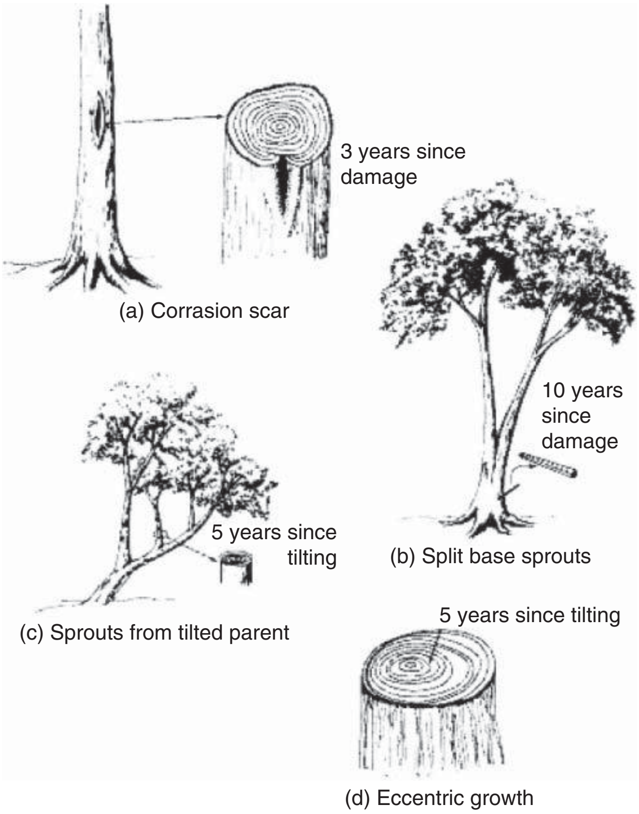


Figure 10.6 Types of botanical evidence of geomorphic processes. Source: Hupp, 1988. Reproduced with permission of Wiley.

Figure 10.7 Vegetation banding on a point bar along the Cecina River, Italy. Note different species and size classes in bands along the extension axis of the bar. People, for scale, on mid-ground on right. Photograph by C.R. Hupp.

of the sprout must be included in the core or cross-section. Some trees may bear evidence of several floods (Fig. 10.5). The accuracy obtained from adventitious sprouts is exceeded only by corrasion scar analysis, which, because of healing, may have limited use in dating older events.

Ring anomalies. Geomorphic events may affect trees without leaving obvious evidence as described above; ring anomalies are perhaps the most pervasive form of botanical evidence of geomorphic events. This may occur as changes in ring width or as various alterations in intra-ring tissue in response to flood impact and or prolonged inundation (Yanosky 1983; St. George and Nielsen 2003; Stoffel and Bollschweiler 2008). Cores from trees must be analysed, usually microscopically, to detect anomalous rings or ring patterns. Eccentric ring patterns (Fig. 10.6d) occur when a tree is tilted from vertical. Slight inclinations can induce eccentric rings without causing the formation of adventitious sprouts. Abrupt tilting of a tree (as is typical of some geomorphic events) results in subsequent rings that are wide on one side of the trunk and narrow on the other. This growth response is typically referred to as compression wood on the down or bottom side of the tilted stem and tension wood on the upper or top side of the stem. When this pattern occurs after concentric ring production, the date of the onset of eccentric growth is usually within 1 year of the event. This line of evidence is particularly useful in areas dominated by conifers (Stoffel and Bollschweiler 2008), which do not typically form adventitious sprouts regardless of degree of inclination.

Tree age. Flood-deposited sediment or flood-scoured areas provide ‘new’ sites for vegetation establishment. These surfaces may occur after low-frequency, high-magnitude events such as crevasse splays after a levee break or on avulsed areas. More frequent processes such as meander extension may create/enlarge point bars (Fig. 10.7), longitudinal bars or aggrading banks, which may be rapidly colonized by certain (typically ruderal) woody species (Everitt 1968; Hupp 1988; Johnson 1994; McKenney *et al.* 1995; Scott *et al.* 1997; Friedman *et al.* 2005; Merigliano *et al.* 2013). The age of trees and shrubs growing on these new surfaces indicates a minimum time since initial deposition or scour (Harper 1912). When all the oldest trees on a geomorphically delineated surface are of nearly the same age (even-aged stand), it is reasonable to assume, at least in humid environments, that the age of the oldest is close to the age of the surface, barring subsequent, unrelated, events such as lumbering or fire (Hupp *et al.* 1987; Stella *et al.* 2011). However, depending on characteristics of the deposit, climate and seed source, a substantial variation in time between deposition and plant establishment may occur (Pierson 2007; Stoffel *et al.* 2008; Stella *et al.* 2011). This technique is useful for obtaining minimum ages for surfaces created by large, exceptionally infrequent (hundreds of years) events. Along actively migrating channels, tree ages can be used to estimate the timing and rate of channel shifts. Point bars (Fig. 10.7), in particular, may have several bands of woody plants whose age is progressively younger towards the channel (Hupp 1988; McKenney *et al.* 1995; Merigliano *et al.* 2013).

Sediment deposition and erosion

Substantial amounts of sediment can be deposited on or eroded from alluvial areas during flooding events. Rates and amount of sediment deposition, scour and associated channel shifting can be estimated using woody vegetation (e.g. Sigafoos 1964; Hupp 1988, 1992, 2000). Suspended sediment is an especially important environmental concern because of problems associated solely with sediment and because of problems associated with the adherence of hydrophobic contaminants to fine particles (Noe and Hupp 2009). Unfortunately, sedimentation trapping rates, fluxes on and off the floodplain and other riparian surfaces, and retention times are poorly understood. Indeed, as of the early 1990s, only four accounts of vertical sediment accretion rates in any type of wetland in the United States had been published (Johnston 1991). Since then, other streams, particularly in the southeastern United States (Hupp 2000; Noe and Hupp 2009; Aust *et al.* 2012) have been studied for sediment trapping rates, but worldwide the number of riparian areas with documented sedimentation rates is low. Woody vegetation analyses offer an inexpensive and relatively accurate method for obtaining net sedimentation rate data along many streams. Further, buried stems may allow for the documentation and timing of episodic depositional events.

Buried stems are the principal form of botanical evidence used to estimate sedimentation rates (Sigafoos 1964; Hupp and Simon 1991; Scott *et al.* 1996). Initial tree roots (at germination) grow just below the ground surface and eventually form the major root trunks that radiate horizontally from the germination point. The basal flare or root collar and initial root zone are thus a distinctive marker of the original ground surface at the time of germination (Fig. 10.8). Trees subjected to substantial sediment deposition typically have the appearance of a telephone pole because burial may obscure the normal flare at the base of the tree (Fig. 10.8). Net sedimentation rates can be estimated by determining the depth of root burial some distance from the exposed trunk (usually one or two tree diameters), then extracting a tree core near the ground surface and determining the age of the tree and finally dividing the depth of burial by the age of the tree (Alestalo 1971; Hupp 1988; Strunk 1997; Ross *et al.* 2004; Piégay *et al.* 2008; Kroes and Hupp 2010). Several trees at a location should be analysed similarly to obtain an average net rate to reduce possible micro-site variation. This technique yields sedimentation rate estimates over the average age of the sampled trees (Hupp and Bazemore 1993; Ross *et al.* 2004; Dufour and Piégay 2008; Provansal *et al.* 2010) for specific areas over a bottomland and has been shown to be usefully accurate compared with repeat cross-sectional analyses (Hupp and Simon 1991; Hupp *et al.* 1993), land use change (Mizugaki *et al.* 2006) and marker horizons (Jolley *et al.* 2010; Kroes and Hupp 2010). However, estimates of short-term (past 1–5 years) sedimentation rates may be measured more accurately using artificial marker horizons such as white feldspar clay pads (Hupp and Bazemore 1993; Kleiss 1996; Kroes and Hupp 2010).

Excavation of buried trees (Figs 10.5 and 10.8) allows access to buried stems where dendrochronological analyses of wood tissue may yield detailed sedimentary histories over the life of the tree. Because woody tissue produced after stem burial closely resembles root tissue, it may be distinguished from previously formed stem wood (above ground) and allow the estimation of timing and measurement of sediment deposits (Sigafoos 1964). Various anatomical changes may occur after burial, including increased vessel size (Nanson and Beach 1977) and abrupt reductions in ring width (Rubtsov and Salmina 1983). A detailed depositional history using both of the above features was constructed by Friedman *et al.* (2005) that included several events. Merigliano *et al.* (2013) provide a detailed summary of the use of this line of botanical evidence.

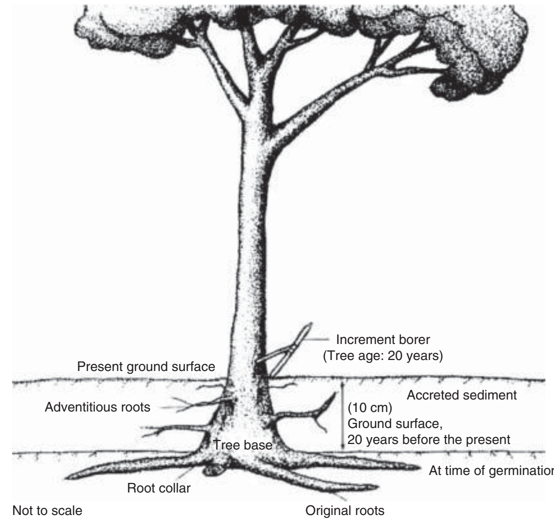
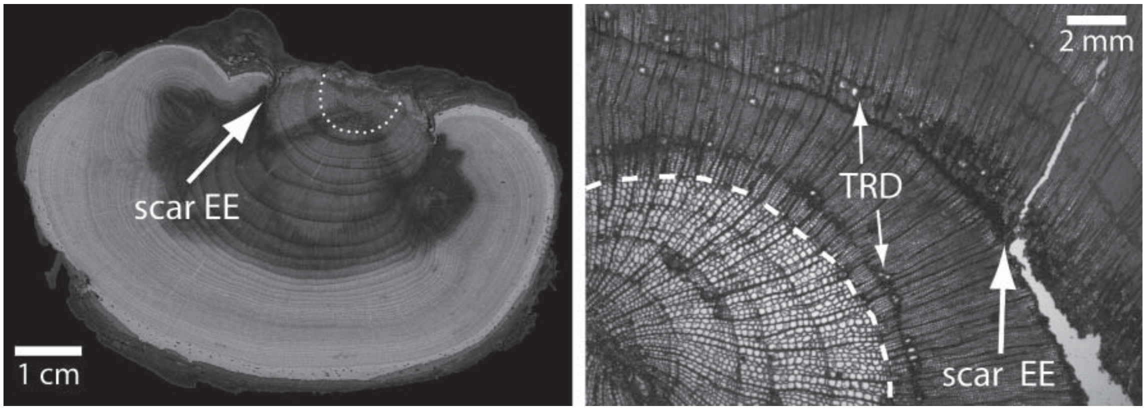


Figure 10.8 Generalized 'buried' floodplain tree. Determination of depth to root collar at time of germination (just below original ground surface) shows a net accretion of 10 cm. An increment core taken at the base of the tree indicates that the tree is 20 years old. Thus, a net sedimentation rate is conservatively estimated to be 0.5 cm per year for the past 20 years. Hupp and Morris, 1990. Reproduced with permission of Springer.

The use of dendrogeomorphic techniques to document erosion rates has largely been confined to hillslope processes (LaMarche 1968; Shroder 1978; Hupp and Carey 1990), including rock slides, avalanches and debris flows (Hupp 1983, 1984; Stoffel *et al.* 2005; Hitz *et al.* 2008). In eroding areas, the lateral roots become exposed. The technique for estimating erosion is similar to that used in estimating deposition; measurements are made from the top of the exposed roots to the present ground surface to obtain a depth (Fig. 10.5). The tree is then cored near its base to determine the age, which is divided into the erosion depth to obtain a net erosion rate over the life span of the tree. Gärtner (2007) reviewed the methods that use root exposure and growth responses to measure erosive processes. Sigafoos (1964) used this technique to document flood-related cycles of deposition and erosion along a large river floodplain. Similarly, rates of bank retreat have been estimated using this technique (Simon and Hupp 1987). Root exposure, like stem burial in reverse, may cause anatomical changes that can be used to date erosive events (Hitz *et al.* 2008). Roots exposed along eroding cut banks or channels impacted by incision have been a tantalizing feature to measure, yet there have been few studies



(a) (b)

Figure 10.9 (a) Cross-section of exposed root (Austrocedrus chilensis) with initial exposure signature, change to stem wood formation in 1956 (changes in cell structure; dashed line) and subsequent corrasion scar (arrow) in early wood (EE) in 1960. (b) Microsection of Douglas fir (Pseudotsuga menziesii) root with initial exposure (dashed line) in 2004, as well as scar (arrow) and traumatic resin ducts (TRD), in 2006 and 2007. Stoffel *et al*, 2012. Reproduced with permission of GSA Journals.

(and only recently, Stotts *et al.* 2013) that have analysed the ring structure of exposed roots in riparian situations. Stoffel *et al.* (2012) provide an innovative approach (Fig. 10.9) to estimate bank retreat using root exposure in the Patagonian Andes. Root exposure on floodplains has also been used to estimate soil surface subsidence (Kroes and Hupp 2010).

Temporal trends

One of the more unique but underutilized applications of dendrogeomorphology is the detection of changes in geomorphic processes, particularly deposition and erosion over time, that may be related to environmental or stream regime shifts. It is possible to infer changes in these processes by organizing the dendrogeomorphic data into tree-age classes (cohorts) over the life span of most of the samples. Intentionally sampling many trees over a wide range of ages allows for the calculation of mean rates of deposition over, say, decadal time periods (Hupp and Bazemore 1993; Heimann and Roell 2000; Piégay *et al.* 2008; Kroes and Hupp 2010). However, there are several caveats that must be assumed: all rates are net rates, compaction is not normally taken into account and the accuracy of early rates of deposition is damped because subsequent deposition affects the calculation of any prior time period. It is possible that some sites may experience higher deposition rates than measured if erosion occurs during intervening periods. Subtracting out subsequent deposition is usually subjective, provides additional error and does not typically improve the overall interpretative ability (Hupp and Bazemore 1993).

Temporal trends in deposition/erosion can be estimated using dendrogeomorphic techniques across single banks (Hupp and Simon 1991; Provansal *et al.* 2010), along extended reaches (Friedman *et al.* 1996; Piégay *et al.* 2008) and over broad floodplain areas (Hupp and Morris 1990; Hupp and Bazemore 1993; Ross *et al.* 2004; Kroes and Hupp 2010). Hupp and Bazemore (1993) found that channelized rivers in West Tennessee consistently through time have lower floodplain sedimentation rates than unchannelized rivers, not unexpected given that channelization intentionally and typically severely reduces overbank flooding. Further, both systems simultaneously responded with increased sedimentation rates after substantial bottomland land clearing for agriculture following the close of World War II. Although the sedimentation rate for a historical decade may be imprecise, it is nevertheless possible to make relative comparisons that may yield valid environmental interpretations. Few other relatively inexpensive techniques may provide long-term sedimentation rates in this detail. For example, in cut-off channels along the high-energy Ain River, France, Piégay *et al.* (2008) found that sedimentation processes were characterized by shifting braided, meandering and wandering reaches. Deposition rates varied according to age of cut-off, channel type, channel movement and within site distances from stream edge (Fig. 10.10).

10.5 Description of fluvial landforms through vegetation

Fluvial geomorphic processes via the action of flowing water create a number of widely recognized landforms, from small channel bedforms to extensive floodplains. Some of the most influential effects include (i) the creation of new areas for establishment such as point bars, depositional islands, abandoned channels and scour pools; (ii) variations in bank stability related to various stream types that may occur under a given climatic condition; (iii) the formation of flood intensity gradients normal to the low-flow channel; (iv) spatial variation in sediment deposition and erosion rate, size clasts, subsequent hydrogeomorphic connectivity with river seepage and aquifers and nutrient availability; and (v) gradients related to surficial connectivity with stream flow and inundation duration (hydroperiod). Separating these and other factors that influence bottomland vegetation patterns is difficult because most are

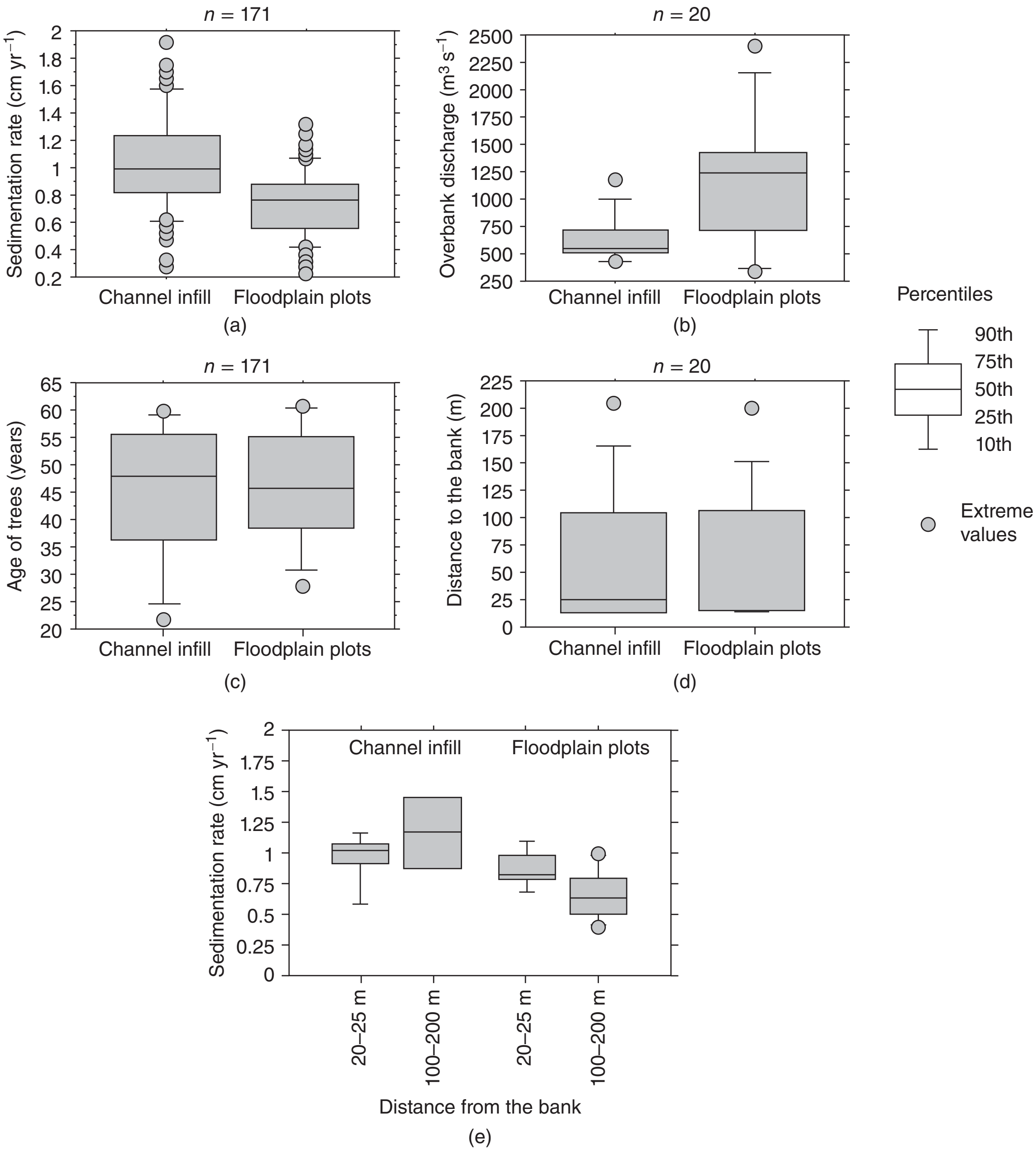


Figure 10.10 Comparison of sedimentation rate and local features between channel infills and the alluvial forest plots. Box plots of (a) the sedimentation rate, (b) the overbank flow discharge, (c) the age of trees, and (d) distance to river bank for infill and floodplain plots. Sedimentation rate variation against distance river bank for infill and floodplain plots (e). Piégay *et al*, 2008. Reproduced with permission of AGU Journals.

distinctly interdependent and a general lack of consistent or accepted landform and process definitions exists within the geomorphic sciences and especially between the geomorphic and ecological sciences. Yet precisely for these reasons there is a need for analyses of vegetation patterns that may ultimately allow for the appropriate discernment of the most proximal geomorphic ecosystem drivers.

Fluvial landforms and floods

Floodplains, like most fluvial landforms, are dynamic features eroding in places while aggrading in others. Channel dynamics, especially those of meandering, pseudo-meandering, wandering and braided streams, provide the energy necessary to erode and transport floodplain sediments. Meanders typically extend, eroding accreted sediments until they are cut off by an avulsion (channel cut-off), leaving an oxbow lake and a new channel. Whole meander loops, additionally, tend to migrate downstream. Thus, over geomorphic time, nearly all floodplain alluvium is in a state of flux. Large rivers that drain Alpine areas and their piedmonts (e.g., Europe and western North America) tend to have relatively high energy, abundant gravel bedload and straight, braided or wandering channel patterns (Bravard *et al.* 1986; Kellerhals and Church 1989; Müller 1995). Lateral instability and avulsion in these systems during even moderate flows promote the development of abandoned channels such as chute cut-offs, oxbow lakes and dead arms (Bravard and Gilvear 1996). This lateral mobility of the main channel increases the biodiversity on these floodplains by creating coincident patches with different ages and hydrological characteristics (Pautou 1984; Amoros *et al.* 1987; Amoros and Roux 1988). Large upland rivers draining relatively moderate gradient areas (e.g., eastern North America) tend to carry considerable silt/clay sediment loads and may be relatively stable with large, fine-grained floodplains. Typically high-magnitude, low-frequency flow events are necessary for significant avulsion activity. However, consistent variation in hydrogeomorphic conditions with surface elevation creates long linear patches analogous to the mosaic on the more dynamic Alpine rivers.

Floodplains of large lowland rivers (e.g. the Coastal Plain of the eastern United States) tend to have net sediment storage during high or rising sea levels such as the conditions over the past several thousand years (Hupp 2000). These lowland rivers, with active meandering, tend to be more laterally dynamic than stable upland rivers with moderate gradients, and share many of the hydrogeomorphic characteristics of the high-energy Alpine streams. Lowland floodplains aggrade in two ways: first, by lateral accretion or point-bar extension, where coarse material is deposited on the inside bank of channel bends; a corresponding volume is typically eroded on the opposite or cut bank; and second, by vertical accretion where suspended sediment (typically fines) is deposited over the floodplain during overbank flows. Lateral accretion is an episodic process that occurs during high flows, building the point bar into an often crescent-shaped ridge. Over time, a series of high-flow events produce the ridge and swale topography associated with meander scrolls. The establishment of ruderal woody vegetation during intervening low-flow periods on fresh scroll surfaces creates bands (Fig. 10.7) of increasingly younger vegetation towards the main channel (McKenney *et al.* 1995). These bands of vegetation may accentuate the ridge and swale topography by creating contrasting depositional environments during subsequent high flows; the hydraulics necessary to produce meander scroll topography and the role of vegetation in its development are poorly understood.

The often drastic and sudden reduction in flow velocity after leaving the main channel and entering the hydraulically rough floodplain environment facilitates fine-sediment deposition. As rising floodwaters overtop the bank, the coarser (or heavier) sediment is deposited first, creating natural levees along the floodplain margin. Levees tend to be most pronounced along relatively straight reaches between meanders and are often the highest ground on the floodplain. Levees are sometimes breached by streamflow resulting in a crevasse splay that may insert coarse material deep into the otherwise fine-grained bottom. Natural levee development and the breaches that form are poorly documented in the literature, yet are critical in the understanding of the surface-water hydrology of most bottomlands and present an area ripe for dendrogeomorphic study. Levee height and breaches strongly affect the hydroperiod (and thus, sedimentation dynamics) in systems dominated by surface-water flow (Patterson *et al.* 1985).

River overflows can result in flooding over large areas of the floodplain that may have only minor erosion or, in the case of spates (flash floods), considerable scouring, massive slope failure and bank erosion may occur (Resh *et al.* 1988; Ward *et al.* 1999). In the latter, which may be more typical of Alpine braided rivers, erosion occurs with aggradation, leading to a shifting mosaic of scoured and aggraded patches, depending on the local slope and coarse material supply (Kalliola *et al.* 1991; Bravard and Gilvear 1996). Similarly, at the scale of river reaches, scouring occurs when sufficient slope is combined with low sediment supply from upstream (Petts and Bravard 1996). Conversely, when the gradient is low and sediment supply is high, aggradation is likely. Non-equilibrium situations, such as chronic incision or aggradation, usually result from anthropogenic impacts (Galay 1983; Bravard *et al.* 1997). Any decrease in coarse material supply or any alteration of the river transport ability can lead to the incision of the riverbed. Channel incision triggers a progressive disconnection of the riparian zone and riverine wetlands from the river and thus to a decrease in scouring frequency (Bornette and Heiler 1994; Bravard *et al.* 1997) and floodplain inundation (Hupp 1999). Models of channel evolution during and after incision with direct references to vegetation dynamics and diversity are presented in Simon and Hupp (1987) and Hupp and Rinaldi (2007).

Reading landforms through vegetation

It is implied in the following discussion of geomorphic forms and processes that affect and maintain vegetation patterns that the vegetation similarly affects the geomorphology. Thus, patterns in both vegetation and geomorphology are intimately related and mutually sustained. Therefore, vegetation patterns may be used as a tool to predict ambient geomorphic conditions. The fluvial landscape is a shifting mosaic of landforms adjacent to stream channels (Bravard *et al.* 1986; Swanson *et al.* 1988) and/or a relatively predictable, largely linear array of landforms (Osterkamp and Hupp 1984; Hupp and Osterkamp 1996). Regardless, considerable variation in hydrogeomorphic processes can occur over short distances across this landscape and indeed across a single fluvial landform such as a floodplain. This complexity of physical form and process is reflected in a wide array of riparian community patterns (Hupp and Osterkamp 1985; Nilsson *et al.* 1989; Naiman *et al.* 1993). Periodic disturbance by relatively frequent floods that may scour or aggrade and/or inundation duration (hydroperiod) have been cited as the principal fluvial geomorphic processes responsible for the high biodiversity in riparian ecosystems (Vannote *et al.* 1980; Hupp and Osterkamp 1985, 1996; Nilsson *et al.* 1989; Gregory 1992; Sharitz and Mitsch 1993; Bornette *et al.* 1994, 2008; Hupp 2000; Nakamura *et al.* 2007) and the maintenance of pioneer habitats in the fluvial corridor (Ward *et al.* 2001; Hughes *et al.* 2005; Van Looy *et al.* 2008).

Common fluvial landforms (Fig. 10.2), as defined by Osterkamp and Hupp (1984), occur as terraces high in the valley section (flow return interval >3 years) and, in descending order, proceed through floodplain (flow return interval every 1–3 years), various riparian features on banks (flow return interval <1 year, usually measured by annual flow duration), channel bars to the channel bed. Studies across continents have shown that there are characteristic plant-species distributional patterns for specific fluvial landforms and processes in temperate biomes (Osterkamp and Hupp 1984; Décamps *et al.* 1988; Gregory 1992; Naiman *et al.* 1993; Pautou and Arens 1994; Marston *et al.* 1995; Hupp and Osterkamp 1996; Bravard *et al.* 1997; Hughes 1997; Tabacchi *et al.* 1998; Bendix and Hupp 2000; Nakamura and Shin 2001; Gurnell and Petts 2003; Tabacchi and Planty-Tabacchi 2005; Steiger *et al.* 2005; Hupp and Rinaldi 2007) and also in other bioclimatic contexts (Hughes 1990; Wyant and Ellis 1990; Kalliola *et al.* 1991; Hoff 1996; Rosales-Godoy *et al.* 1999; Pike and Scatena 2010).

In the fluvial environment, disturbance characteristics play a major role in the development of many vegetation patterns (Johnson *et al.* 1985; Day *et al.* 1988; Kirkman and Sharitz 1994; Bornette *et al.* 2008). Periodic flood disturbances control certain communities that persist in dynamic equilibrium (Pickett 1980; Hupp and Osterkamp 1985; Bornette and Amoros 1991; 1996; Bendix and Hupp 2000) with no loss of species compositional integrity over time. Along low-gradient systems, with extensive forested wetlands, the tight relation between vegetation type and annual length of inundation (hydroperiod) is well documented (Wharton *et al.* 1982; Sharitz and Mitsch 1993). Moderate gradient streams of temperate eastern North America typically develop consistent and persistent linear fluvial landforms that are maintained by predictable variation in discharge. Coincident hydrogeomorphic analyses (Osterkamp and Hupp 1984; Hupp and Osterkamp 1985; Bendix and Hupp 2000) along several of these streams suggest that the overriding influence on the distributional patterns of species is the frequency of inundation and the susceptibility of plants to destructive flooding. Thus, the hydrogeomorphic processes operating differently on the different landforms affect the plant patterns, not the landforms per se. As examples, two shrubs common on the channel shelf, along high-gradient Passage Creek, Virginia (bank feature), *Alnus serrulata* and *Cornus amomum*, are relatively resistant to destruction by floods because of small, highly resilient stems and the ability to sprout rapidly from damaged stumps. Conversely, *Cornus florida* and some species of *Quercus* and *Carya*, which commonly grow on terraces but rarely on lower surfaces, may be intolerant of repeated flood damage or inundation. Floodplain species, such as *Carya cordiformis*, *Juglans nigra* and *Ulmus americana*, are less tolerant of destructive (scouring) flooding than are channel-shelf species, but more tolerant of periodic inundation than are terrace species. Depositional bars (except point bars) rarely support woody species; however, several perennial herbaceous species survive destruction through deeply anchored perenating rootstocks.

Across most low-gradient lowland floodplains, striking vegetation zonation is displayed (e.g., Coastal Plain of the southeastern United States). Small differences in elevation, often measured in centimetres, may lead to pronounced differences in length of inundation (hydroperiod) and thus community composition (Mitsch and Gosselink 1993). Many vegetation classification systems infer that hydroperiod is the most influential factor affecting species patterns (Fig. 10.11). It should be noted that flooding, per se, may not only limit species distribution but also the long-lasting light deficiency in submergent vegetation and anaerobic conditions associated with flooding (Wharton *et al.* 1982). Plants tolerant of varying degrees of flooding have developed physical and/or metabolic adaptations to deal with inundation and anoxia (Wharton *et al.* 1982; Vartapetian and Jackson 1997; Bornette *et al.* 2008). Presumably it is the degree to which individual species have adapted to anoxia-related stresses that has led to the distinct and drastic changes in vegetation composition over very short distances (metres) across many lowland floodplains (Huffman and Forsyth 1981). Thus, where the vegetation and inundation patterns are closely related, vegetation may be used to infer inundation frequencies in non-monitored parts of the floodplain (Wohl and Merritt 2008; Pike and Scatena 2010).

10.6 Communities as an indicator of disturbance regime

Significant disturbance in riparian areas typically comes from events at both ends of a geomorphic process gradient that has severe erosion at one end and severe aggradation at the other (Fig. 10.1). The vegetation patterns described in this section may be used as an indicator of the level and type of persistent geomorphic disturbance. Aggradation along stream reaches, natural or otherwise, provides new sites for vegetation colonization and may initiate a succession of vegetation stages indicative of progressively changing hydrogeomorphic conditions (Hupp 1992; McKenney *et al.* 1995; Corenblit *et al.* 2011). Only a few species, capable of rapid root growth along newly buried stems (e.g., species of *Salix* and *Populus*), may occupy bottomland areas periodically affected by frequent or large amounts of sediment deposition (Fig. 10.10). Conversely, erosion and channel cut-offs may lead to nearly permanent ponded areas that support aquatic vegetation. Hydraulic scour and size-clast sorting are important factors for plant colonization and survival. Obviously, where erosion removes all or part of the soil in the root zone of trees, they will be killed or damaged (Fig. 10.9). Initial vegetation establishment increases hydraulic roughness, facilitating further sedimentation, and ultimately modifies the relatively unstable initial surfaces into relatively stable surfaces favourable for the recruitment of later, stable-site species (Hupp

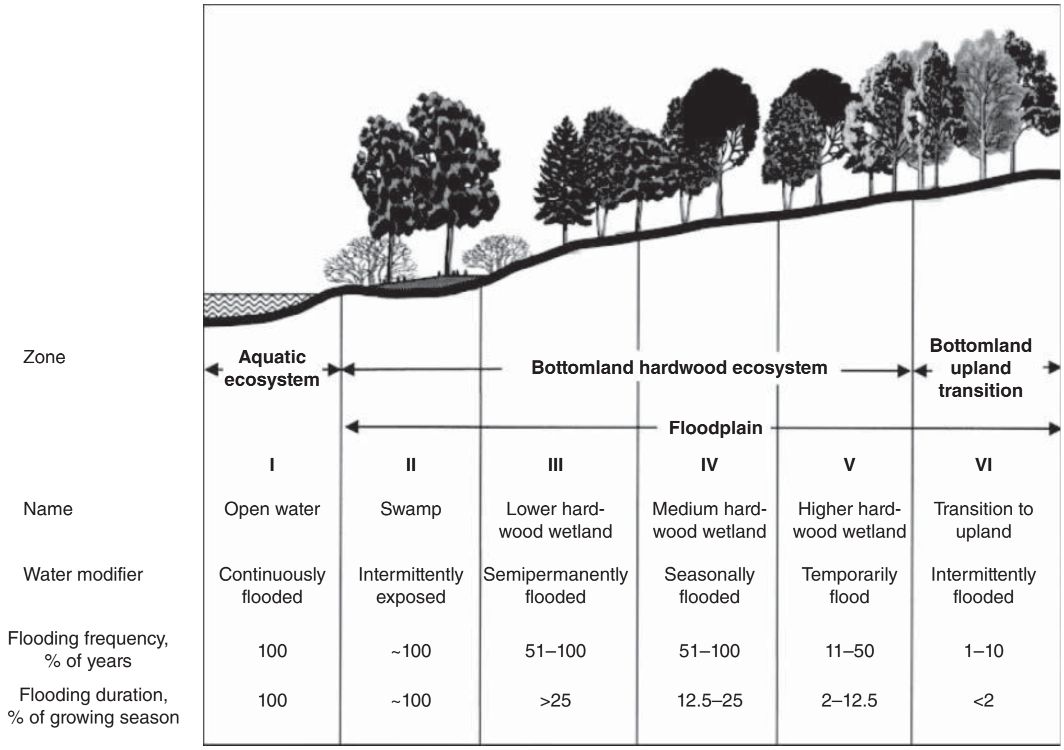


Figure 10.11 Zonal classification of bottomland hardwood ecosystems (lowland floodplain/terrace, southeastern United States) showing average hydrological conditions for each vegetatively distinct zone. Shartiz and Mitsch, 1993. Reproduced with permission of Wiley.

1992; Johnson 1994; Friedman *et al.* 1996b; Corenblit *et al.* 2007). In all situations, ruderal species having life history characteristics including abundant dependable seed production or vegetative propagule crop, effectively dispersed by wind or water to suitable riparian sites, rapid recruitment, shoot and root development are usually favoured in flooded areas (Hupp 1992; Johnson 1994; Karrenberg *et al.* 2003; Bornette *et al.* 2008).

Along streams dominated by aggradation, extensive depositional areas may be generated during each flood. Such floodplains are common, for example, along many southeastern US streams where upstream agriculture practices and channelization generate large amounts of suspended fines (Bazemore *et al.* 1991; Hupp 1992). In the upstream degradation situation, large eroded zones are generated and the river is decreasingly connected to its floodplain with some important consequences for riparian species and communities (Bravard *et al.* 1997; Wyzga 1999; Steiger *et al.* 2001; Dufour and Piégay 2008). Such conditions (e.g., reduction in water-table elevation) are common in rivers subjected to incision, for example, as a consequence of flow regulation (Galay 1983; Babinsky 1992; Bravard 1994; Peiry *et al.* 1994; Hupp 1999). Most streams in the West Tennessee portion of the southeastern Coastal Plain of the United States have been channelized (Simon and Hupp 1992). This channelization led to severe degradation or erosion of the affected and upstream reaches of the streams. Degradation occurs on both the channel bed and banks until some quasi-equilibrium is attained and aggradation begins low in the channel section (Simon and Hupp 1992). A cycle of erosion, accretion and return to equilibrium is described in a six-stage model that incorporates vegetation in the geomorphic processes (Simon and Hupp 1987, 1992; Hupp 1992). Similarly, reaches below high dams may also experience severe incision/erosion (Williams and Wolman 1984; Bravard *et al.* 1997; Friedman *et al.* 1998; Brandt 2000) and follow similar geomorphic responses as described in the model above (Hupp *et al.* 2009) with analogous vegetation succession described below. Over half of the world's largest river systems (172 of 292) have been moderately to strongly affected by dams (Nilsson *et al.* 2005).

The erosional phase of this channel evolution model often completely removes all woody bank vegetation. Late in this phase, refugia occur in protected areas, usually downstream of slump blocks from mass wasting on the banks (Simon and Hupp 1992). These refugia offer enough stability for ruderal riparian vegetation to establish. However, upland species may occur high on the banks and represent species especially tolerant of erosive conditions. During the highly aggradational phase of this cycle, dominant species (found in refugia of the late erosional phase) are tolerant of high sediment accretion rates. The equilibrium phase of this cycle is, by far, the most diverse and is always vegetated. Early equilibrium sites may support high percentages of some species most indicative of aggradation, which experience substantial declines in percentage occurrence as sites regain equilibrium. Many species normally found in non-impacted bottomland hardwood forests are present in the equilibrium phase.

Similarly, in aquatic ecosystems, scouring, in addition to flooding, affects vegetation patterns. Scouring during river overflow is considered a disturbance that may uproot plants and disrupt communities (Jones 1956; van der Valk and Bliss 1971; Bilby 1977) and in some cases, may completely remove fine sediment deposits that had accumulated since the last scouring event. Natural successional processes that occur in riverine ecosystems can be slowed or stopped, depending on the intensity and frequency of flood scouring (Sparks *et al.* 1990; Foeckler *et al.* 1991, 1994; Müller 1995; Bornette and Amoros 1996). In accordance with the intermediate disturbance hypothesis (Connell 1978; Sousa 1984), intermediate frequency and intensity of flood scouring sustain a dynamic equilibrium at least in some Alpine European ecosystems. It allows for the maintenance of a highly diverse, shifting mosaic of plant species (Resh *et al.* 1988; Bornette and Amoros 1991; Roberts and Ludwig 1991; Bornette *et al.* 1998). Along the Ain River, France, in the piedmont of the Jura mountains, high flood scouring frequency and/or intensity usually leads to a decrease in biodiversity, because plant communities are unable to recover before the next disturbance or because of a lack of substrate favourable for establishment (Kohler and Schiele 1985; Resh *et al.* 1988). Conversely, low-frequency and/or low-intensity flood scouring is unable to impede completely successional processes and substrate accumulation that may occur, leading to the terrestrialization of aquatic ecosystems and a progressive reduction in flood frequency. Large-sized plants (*Nuphar lutea*, *Nymphea alba*) unable to resprout from above ground parts typically colonize undisturbed or rarely disturbed channels (Bornette *et al.* 1998; Amoros and Bornette 1999). Conversely, generally moderately scoured habitats, maintained at a dynamic equilibrium stage by disturbance frequency/intensity, typically support highly diverse communities of evergreen perennial species (Greulich and Bornette 1999) able to resprout efficiently from any broken part of their individual (Barrat-Segretain *et al.* 1998, 1999). In lowland floodplains or in locally aggradational areas, rooted plants may ultimately be eliminated and replaced by unanchored plants (*Ceratophyllum demersum*, Lemnaceae), but moderate silting may allow the persistence of some species (such as *Vallisneria* sp., *Najas* sp.) (Haslam 1978; Rybicki and Carter 1986; Amoros *et al.* 2000).

Floodplain submersion during floods is a major limiting factor for aquatic vegetation, particularly for long-duration (Fig. 10.11) or high-discharge floods. Sediment deposition usually increases the terrestrialization rate not only through increasing aggradation, but also through eutrophication. Commonly, in aggrading permanently connected waterbodies, floating plants are easily washed away even by water with low velocity (Bornette *et al.* 1998). Hence both species able to resprout after burial and annual fast-growing species are usually favoured in such situations (Haslam 1978; Kalliola *et al.* 1991; Amoros and Bornette 1999) and their presence may be used to infer ambient flooding conditions. Ultimately, sediment deposition may eliminate aquatic plants from parts of the riparian ecosystem.

10.7 Conclusions

Fluvial geomorphic processes and landforms exert a profound influence on individual plants and vegetation patterns. Thus, inference may be drawn on hydrogeomorphic conditions through the analysis of dendrogeomorphic evidence, riparian species composition, diversity and life-history characteristics. Vegetation may be used as a geomorphic tool in three basic ways. First, vegetation may inform general fluvial geomorphic characters or processes acting at the reach scale, mainly in a qualitative or semiquantitative way (Table 10.1). Second, vegetation may also quantify more specific geomorphic processes such as sedimentation/erosion rates and/or landform age. Third, vegetation helps to quantify or characterize hydrological processes or parameters very useful for geomorphic studies. Among these tools, dendrogeomorphology (tree-ring dating) has a very important place in terms of frequency and variety of uses (see Table 10.1). It provides quantitative information in a wide range of fluvial applications, including floods, floodplain deposition and aggradation, channel dynamics and mobility, estuary and lake shoreline dynamics, saltwater intrusion along streams and mountain glacier activity and debris flows. Where historic records are short or lacking, tree-ring study may be the most accurate method for obtaining magnitude and frequency data over the past few hundred years. Botanical evidence, as a tool, in combination with geomorphic evidence allows for the interpretation and determination of the relative importance of various geomorphic processes.

Predictable patterns of species composition and community structure occur along high-energy Alpine rivers, medium-energy stable piedmont rivers and low-energy lowland rivers. Also, a similar pattern may occur along a flow regime gradient and the vegetation response may be directly related to specific hydrogeomorphic conditions. Vegetation organization, composition and plant community dynamics on river floodplains are controlled by (i) disturbance type (erosion, deposition) and scale (frequency and intensity of erosion vs. deposition and flood duration) and (ii) biological characteristics of plants linked to resistance to disturbance, resilience (diaspore type, abundance and dispersability, recruitment efficiency) and competitive ability. To understand fluvial systems fully, one needs to study both the plants and the substrate and a complete understanding of each can only be gained by studying the other. In particular, studies of plants may elucidate more about geomorphology than the study of geomorphology alone.

Although a huge amount of ecogeomorphic research shows that vegetation can successfully be used as an indicator, there are some considerations that, in certain situations, limit its use as an efficient tool (Dziock *et al.* 2006). First, if many authors show that physical drivers have some direct impacts on vegetation (community composition, diversity, growth) (e.g. Dufour and Piégay 2008; Jolley *et al.* 2010), one vegetation indicator (a species presence, for example) may require a combination of several drivers for explanation because of the complexity that the vegetation/fluvial dynamic links made with non-linear processes, thresholds, lag effects and nested drivers. Some vegetation patterns often result from various drivers, hence the link between vegetation and fluvial process or pattern is not always an univocal link (Willms *et al.* 1998; Baker and Wiley 2004; Merigliano 2005; Dufour *et al.* 2007; Dufour and Piégay 2010; Lowe *et al.* 2010; Sambaré *et al.* 2011) and some parts of the variability in vegetation patterns remain unexplained by geomorphological or hydrological drivers (Budke *et al.* 2010; Angiolini *et al.* 2011). Additionally, a similar biotic response may arise from apparently distinct hydrogeomorphic contexts (Baker and Wiley 2004). Some specific uses of vegetative tools can be geographically or temporally limited (Willms *et al.* 1998). For example, tree-ring dating techniques obviously require climates with regular patterns in temperature and/or moisture that force periodic dormancy, which leads to discrete tree growth periods. However, the recent emphasis on ecogeomorphic studies will probably increase our ability to provide useful and standardized tools for fluvial studies.

References

Alestalo, J. 1971. Dendrochronological interpretation of geomorphic process. *Fennia* 105: 1–140.

Amoros, C. and Bornette, G. 1999. Antagonist and cumulative effects of connectivity: a predictive model based on aquatic vegetation in riverine wetlands. *Archiv für Hydrobiologie, Supplement* 115(3): 311–327.

Amoros, C. and Roux, A.L. 1988. Interactions between water bodies within the floodplains of large rivers: function and development of connectivity. In: Schreiber, K.F., ed. *Connectivity in Landscape Ecology*, vol. 29, Münstersche Geographische Arbeiten, Münster: Universität Münster, Institut für Geographie, pp. 125–130.

Amoros, C., Rostan, J.C., Pautou, G. and Bravard, J.P. 1987. The reversible process concept applied to the environmental management of large rivers systems. *Environmental Management* 11: 607–617.

Amoros, C., Bornette, G. and Henry, C.P. 2000. A vegetation-based method for the ecological diagnosis of riverine wetlands. *Environmental Management* 25: 211–227.

Angiolini, C., Nucci, A., Frignani, F. and Landi, M. 2011. Using multivariate analyses to assess effects of fluvial type on plant species distribution in a Mediterranean river. *Wetlands* 31: 167–177.

Astrade L. 2005. *La Saône en Crue: Dynamique d'un Hydrosystème Anthropisé*, Lyon: Presses Universitaire de Lyon, 268pp.

Astrade, L. and Begin, Y. 1997. Tree-ring response of *Populus tremula* L. and *Quercus robur* L. to recent spring floods of the Saone River, France. *Ecoscience* 4: 232–239.

Aust, W. McKee, S., Sieler, J., Stram, B. and Schilling, E. 2012. Long-term sediment accretion in Bottomland Hardwoods following timber harvest disturbances in the Mobile-Tensaw River Delta, Alabama, USA. *Wetlands* 32: 871–884.

Babinsky, Z. 1992. Hydromorphological consequences of regulating the Lower Vistula, Poland. *Regulated Rivers – Research and Management* 7: 337–348.

Baker, M.E. and Wiley, M.J. 2004. Characterization of woody species distribution in riparian forests of lower Michigan, USA using map-based models. *Wetlands* 24: 550–561.

Ballesteros Cánovas, J.A., Eguibar, M., Bodoque, J.M., Díez-Herrero, A., Stoffel, M. and Gutiérrez-Pérez, I. 2010. Estimating flash flood discharge in an ungauged mountain catchment with 2D hydraulic models and dendrogeomorphic palaeostage indicators. *Hydrological Processes* 25: 970–979.

Barrat-Segretain, M.H., Bornette, G. and Vilas-Boas, A. 1998. Comparative abilities of vegetative regeneration of aquatic plants growing in disturbed habitats. *Aquatic Botany* 60: 201–211.

Barrat-Segretain, M.H., Henry, C. and Bornette, G. 1999. Regeneration and colonization of aquatic plant fragments in relation with the disturbance frequency of their habitats. *Archiv für Hydrobiologie* 145: 111–127.

Bazemore, D.E., Hupp, C.R. and Diehl, T.M. 1991. *Wetland Sediment Deposition and Vegetation Patterns Near Selected Highway Crossings in West Tennessee*, US Geological Survey Water Resources Investigations Report 91-4106, Nashville, TN: US Geological Survey, 46pp.

Beechie, T. J., Liermann, M., Pollock, M.M., Baker, S. and Davies, J. 2006. Channel pattern and river-floodplain dynamics in forested mountain river systems. *Geomorphology* 78: 124–141.

Bégin, Y. and Filion, L. 1995. A recent downward expansion of shoreline shrubs at Lake Bienville (subarctic Quebec). *Geomorphology* 13: 271–282.

Bégin, Y., Langlais, D. and Cournoyer, L. 1991. Tree-ring dating of shore erosion events (Upper St. Lawrence stream, eastern Canada). *Geografiska Annaler* 73A: 53–59.

Belletti, B., Dufour, S. and Piégay, H. 2015. What is the relative effect of space and time to explain the braided river width and island pattern at a regional scale? *River Research and Applications* 31: 1–15.

Bendix, J. and Hupp, C.R. 2000. Hydrologic and geomorphic impacts on riparian plant communities. Special Issue. *Hydrologic Processes* 14: 2977–2990.

Bennett, S. and Simon, A. 2004. *Riparian Vegetation and Fluvial Geomorphology, Water Science and Application 8*, Washington, DC: American Geophysical Union, 282pp.

Bilby, R. 1977. Effects of a spate on the macrophyte vegetation of a stream pool. *Hydrobiologia* 56: 109–112.

Bollschweiler, M. and Stoffel, M. 2010. Changes and trends in debris-flow frequency since AD 1850: results from the Swiss Alps. *The Holocene* 20: 907–916.

Bollschweiler, M., Stoffel, M., Ehmisch, M. and Monbaron, M. 2007. Reconstructing spatio-temporal patterns of debris-flow activity using dendrogeomorphological methods. *Geomorphology* 87: 337–351.

Bornette, G. and Amoros, C. 1991. Aquatic vegetation and hydrology of a braided river floodplain. *Journal of Vegetable Science* 2: 497–512.

Bornette, G. and Amoros, C. 1996. Disturbance regimes and vegetation dynamics: role of floods in riverine wetlands. *Journal of Vegetable Science* 7: 615–622.

Bornette, G. and Arens, M.F. 2002. Charophyte communities in cut-off channels: the role of river connectivity. *Aquatic Botany* 73: 149–162.

Bornette, G. and Heiler, G. 1994. Environmental and biological responses of former channels to river incision: a diachronic study on the Upper Rhône River. *Regulated Rivers – Research and Management* 9: 79–92.

Bornette, G., Amoros, C. and Chessel, D. 1994. Effect of allogenic processes on successional rates in former river channels. *Journal of Vegetable Science* 5: 237–246.

Bornette, G., Amoros, C. and Rostan, J.C 1996. River incision and decennial vegetation dynamics in cut-off channels. *Aquatic Sciences* 58(1): 31–51.

Bornette, G., Amoros, C. and Lamouroux, N. 1998. Aquatic plant diversity in riverine wetlands: the role of connectivity. *Freshwater Biology* 39: 267–283.

Bornette, G., Tabacchi, E., Hupp, C., Puijalon, S. and Rostan, J.C. 2008. A model of plant strategies in fluvial hydrosystems: *Freshwater Biology* 53: 1692–1705.

Boucher, E., Bégin, Y. and Arseneault, D. 2009. Hydro-climatic analysis of mechanical break ups reconstructed from tree- rings, Necopastic Watershed, northern Québec, Canada. *Journal of Hydrology* 375: 373–382.

Boucher, E., Ouarda, T.B.M.J., Bégin, Y. and Nicault, A. 2011. Spring flood reconstruction from continuous and discrete tree ring series. *Water Resources Research* 47: W07516.

Brandt, S. A. 2000. Classification of geomorphological effects downstream of dams. *Catena* 40: 375–401.

Bravard, J.P. 1994. L'incision des lits fluviaux: du phénomène morphodynamique naturel et réversible aux impacts irréversibles. *Revue de Géographie de Lyon* 69: 5–10.

Bravard, J.P. and Gilvear, D. J. 1996. Hydrological and geomophological structure of hydrosystems. In: Petts, G. and Amoros, C., eds., *Fluvial Hydrosystems*, London: Chapman & Hall, pp. 98–116.

Bravard, J.P., Amoros, C. and Pautou, G. 1986. Impact of civil engineering works on the successions of communities in a fluvial system. *Oikos* 47: 92–111.

Bravard, J.P., Amoros, C., Pautou, G., Bornette, G., Bournaud, M., Creuzé des Châtelliers, M., Gibert, J., Peiry, J.L., Perrin, J.F. and Tachet, H. 1997. Stream incision in Southeast France: morphological phenomena and impacts upon biocenoses. *Regulated Rivers – Research and Management* 13: 75–90.

Brown, A.G. 2002. Learning from the past: palaeohydrology and paleoecology. *Freshwater Biology* 47: 817–829.

Budke, J.C., Jarenkow, J.A. and de Oliveira-Filho, A.T. 2010. Intermediary disturbance increases tree diversityin riverine forest of southern Brazil. *Biodiversity Conservation* 19: 2371–2387.

Butler, D.R. and Hupp, C.R. 2013. The role of biota in geomorphology: ecogeomorphology. In: Shroder, J., Butler, D.R. and Hupp, C.R., eds., *Treatise on Geomorphology, Vol. 12, Ecologeomorphology*, San Diego, CA: Academic Press, pp. 1–5.

Church, M. 1992. Channel morphology and typology. In: Callow, C. and Petts, G., eds., *The Rivers Handbook: Hydrological and Ecological Principles*, Oxford: Blackwell, pp. 126–143.

Cleaveland, M.K. 1980. Dating tree rings in Eastern United States. In: *Proceedings, Dendrology in the Eastern Deciduous Forest Biome, Publication FWS-2-80*, Blacksburg, VA: Virginia Polytechnic Institution and State University, pp. 110–124.

Cleaveland, M.K. 2000. A 963-year reconstruction of summer (JJA) streamflow in the White River Arkansas, USA, from tree-rings. *The Holocene* 10: 33–41.

Cleaveland, M.K. and Stahle, D. 1989. Tree ring analysis of surplus and deficit runoff in the White River, Arkansas. *Water Resources Research* 25: 1391–1401.

Connell, J.H. 1978. Diversity in tropical rain forests and coral reefs. *Science* 199: 1302–1310.

Corenblit, D. and Steiger, J. 2009. Vegetation as a major conductor of geomorphic changes on the Earth surface: toward evolutionary geomorphology. *Earth Surface Processes and Landforms* 34: 891–896.

Corenblit, D., Tabacchi, E., Steiger, J. and Gurnell, A. 2007. Reciproal interactions and adjustments between fluvial landforms and vegetation dynamics in river corridors: a review of complementary approaches. *Earth Science Reviews* 84: 56–86.

Corenblit, D., Gurnell, A.M., Steiger, J. and Tabacchi, E. 2008. Reciprocal adjustments between landforms and living organisms: extended geomorphic evolutionary insights. *Catena* 73: 261–273.

Corenblit, D., Baas, A.C.W., Bornette, G., Darrozes, J., Delmotte, S., Francis, R.A., Gurnell, A.M., Julian, F., Naiman, R.J. and Steiger J. 2011. Feedbacks between geomorphology and biota controlling Earth surface processes and landforms: a review of foundation concepts and current understandings. *Earth Science Reviews* 106: 307–331.

Cowles, H.C. 1901. The physiographic ecology of Chicago and vicinity: a study of the origin, development and classification of plant societies. *Botanical Gazette (Chicago)* 31: 73–182.

Day, R.T., Keddy, P.A. and McNeill, J. 1988. Fertility and disturbance gradients: a summary model for riverine marsh vegetation. *Ecology* 69: 1044–1054.

Décamps, H., Fortuné, M., Gazelle, F. and Pautou, G. 1988. Historical influence of man on the riparian dynamics of a fluvial landscape. *Landscape Ecology* 1: 163–173.

Dick, B.M., Hey, R., Peralta, P., Jewell, I., Simon, P. and Peszlen, I. 2014. Estimating annual riverbank erosion rates – dendrogeomorphic method. *River Research and Applications* 30: 845–856.

Dobrowolski, R., Pidek, I.A., Golub, S. and Dzienkowski, T. 2010. Environmental changes and human impact on Holocene evolution of the Horodyska River valley (Lublin upland, east Poland). *Geochronometria* 35: 35–47.

Dufour, S. and Piégay, H. 2008. Geomorphological controls of Fraxinus excelsior growth and regeneration in floodplain forests. *Ecology* 89: 205–215.

Dufour, S. and Piégay, H. 2010. Channel mobility, hydro-geomorphic disturbances and understory vegetation in floodplain forests (Ain River, France). *Géomorphologie: Relief, Processus, Environnement* 4: 371–386.

Dufour, S., Barsoum, N., Muller, E. and Piégay, H. 2007. Effects of channel confinement on pioneer woody plant community structure, composition and diversity along the River Drôme (SE, France). *Earth Surfaces Processes and Landforms* 32: 1244–1256.

Dufour, S., Muller, E., Straastma, M. and Corgne, S. 2012. Image utilization for the study and management of riparian vegetation: overview and applications. In: Carbonneau, P. and Piégay, H., eds., *Fluvial Remote Sensing for Science and Management*, Chichester: John Wiley & Sons, pp. 215–239.

Dziock, F., Henle, K., Foeckler, F., Follner, K. and Scholz, M. 2006. Biological indicator systems in floodplains – a review. *International Review of Hydrobiology* 91: 271–291.

Eichhorn, B., Neumann, K. and Garnier, A. 2010. Seed phytoliths in West African Commelinaceae and their potential for palaeoecological studies. *Palaeogeography, Palaeoclimatology, Palaeoecology* 298: 300–310.

Everitt, B.L. 1968. Use of cottonwood in an investigation of the recent history of a floodplain. *American Journal of Science* 266: 417–439.

Fagot, P., Gadiolet, P., Magne, M. and Bravard, J.P. 1989. Une étude dendrochronologique dans le lit majeur de l'Ain: la forêt alluviale comme descripteur d'une metamorphose fluviale. *Revue de Géographie de Lyon* 64: 213–223.

Foeckler, F., Diepolder, U. and Deichner, O. 1991. Water mollusk communities and bioindication of Lower Salzach floodplain waters. *Regulated Rivers – Research and Management* 6: 301–312.

Foeckler, F., Kretschmer, W., Deichner, O. and Schmidt, H. 1994. Les communautés de macroinvertébrés dans les chenaux abandonnés par une rivière en cours d'incision, la basse Salzach (Bavière, Allemagne). *Revue de Géographie de Lyon* 69: 31–40.

Follner, K. and Henle, K. 2006. The performance of plants, molluscs and carabid beetles as indicators of hydrological conditions in floodplain grasslands. *International Review of Hydrobiology* 9: 364–379.

Friedman, J.M. and Lee, V.J. 2002. Extreme floods, channel change and riparian forests along ephemeral streams. *Ecological Monographs* 72: 409–425.

Friedman, J.M., Osterkamp, W.R. and Lewis, W.M. 1996a. The role of vegetation and bed-level fluctuations in the process of channel narrowing. *Geomorphology* 14: 341–351.

Friedman, J.M., Osterkamp, W.R. and Lewis, WM. 1996b. Channel narrowing and vegetation development following a great plains flood. *Ecology* 77: 2167–2181.

Friedman, J. M., Osterkamp, W. R., Scott, M. L. and Auble, G. T. 1998. Downstream effects of dams on channel geometry and bottomland vegetation: regional patterns in the Great Plains. *Wetlands* 18: 619–633.

Friedman, J.M., Vincent, K.R. and Shafroth, P.B. 2005. Dating floodplain sediments using tree-ring response to burial. *Earth Surface Processes and Landforms* 30: 1077–1091.

Galay, V.J. 1983. Causes of river bed degradation. *Water Resource Research* 19: 1057–1090.

Garnier A., Neumann K., Eichorn B. and Lespez L. 2013. Phytolith taphonomy in the middle- to late-Holocene fluvial sediments of Ounjougou (Mali, West Africa). *The Holocene* 23: 415–430.

Gärtner, H. 2007. Methodological review and new developments in dating and quantifying erosive processes. *Geomorphology* 86: 243–251.

Gillson, L. and Ekblom, A. 2009. Untangling anthropogenic and climatic influence on riverine forest in the Kruger National Park, South Africa. *Vegetation History and Archaeobotany* 18: 171–185.

Gob, F., Petit, F., Bravard, J.P., Ozer, A. and Gon A. 2003. Lichenometric application to historical and subrecent dynamics and sediment transport of a Corsican stream (Figarella River –France). *Quaternary Science Reviews* 22: 2111–2124.

González Loyarte, M.M. 2003. Relationship among grain-size, plant communities and fluvial and eolian processes in a piedmont of the central Andes in Argentina. *Ecología Austral* 13(1): 27–48.

Grams, P.E. and Schmidt, J.C. 2002. Stream flow regulation and multi-level flood plain formation: channel narrowing on the aggrading Green River in the eastern Uinta Mountains, Colorado and Utah. *Geomorphology* 44: 337–360.

Gregory, K.J. 1976. Lichens and the determination of river channel capacity. *Earth Surface Processes and Landforms* 1: 273–285.

Gregory, K.J. 1992. Vegetation and river channel processes. In: Boon, P.J., Calow, P. and Petts, G.E., eds., *River Conservation and Management*, Chichester: John Wiley & Sons, pp. 255–269.

Gregory, K.J. 2006. The human role in changing river channels. *Geomorphology* 79: 172–191.

Gregory, S., Boyer, K.L. and Gurnell, A.M., eds. 2003. *The Ecology and Management of Wood in World Rivers. American Fisheries Society Symposium 37*, Bethesda, MD: American Fisheries Society.

Greulich, S. and Bornette, G. 1999. Competitive abilities and related strategies in four aquatic plant species from an intermediately disturbed habitat. *Freshwater Biology* 41: 493–506.

Grubb, P.J. 1977. The maintenance of species-richness in plant communities: the importance of the regeneration niche. *Biological Reviews* 52: 107–145.

Gurnell, A.M., and Gregory, K.J. 1995. Interactions between semi-natural vegetation and hydrogeomorphic processes. *Geomorphology* 13: 49–69.

Gurnell, A.M. and Petts, G.E. 2003. Island dominated landscapes of large floodplain rivers: a European perspective. *Freshwater Biology* 47: 581–600.

Gurnell, A.M., Petts, G.E., Hannah, D.M., Smith, B.P.G., Edwards, P.J., Kollmann, J., Ward, J.V. and Tockner, K. 2001. Riparian vegetation and island formation along the gravel-bed Fiume Tagliamento, Italy. *Earth Surface Processes and Landforms* 26: 31–62.

Hack, J.T. and Goodlett, J.C. 1960. *Geomorphology and Forest Ecology of a Mountain Region in the Central Appalachians, US Geological Survey Professional Paper 347*, Washington, DC: US Government Printing Office.

Harper, R.M. 1912. Botanical evidence of the age of certain ox-bow lakes. *Science* 36: 760–761.

Haslam, S.M. 1978. *River Plants*, Cambridge: Cambridge University Press.

Harvey, A.M., Alexander, R.W. and James, P.A. 1984. Lichens, soil development and the age of Holocene valley floor landforms: Howgill Fells, Cumbria. *Geografiska Annaler* 66A(4): 353–366.

Helley, E.J. and LaMarche, V.C. Jr. 1973. *Historic Flood Information for Northern California Streams from Geological and Botanical Evidence, US Geological Survey Professional Paper 485-E*, Washington, DC: US Government Printing Office.

Heimann, D. and Roell, M. 2000. Sediment loads and accumulation in a small riparian wetland system in northern Missouri. *Wetlands* 20(2): 219–231.

Hession, W.C., Wynn, T., Resler, L. and Curran, J. 2010. Geomorphology and Vegetation: interactions, dependencies and feedback loops. *Special Issue. Geomorphology* 116: 203–372.

Hickin, E.J. 1984. Vegetation and river channel dynamics. *The Canadian Geoapher* 28(2): 111–126.

Hitz, O.M., Gärtner, H., Heinrich, I. and Monbaron, M. 2008. First time application of ash (Fraxinus excelsior L.) roots to determine erosion rates in mountain torrents. *Catena* 72: 248–258.

Hoff, M. 1996. Les berges du fleuve Sinnamary (Guyane Française): géomorphologie et groupements végétaux. *Hydroécologie Appliquée* 7(1–2): 151–183 (in French).

Huffman, R.R. and Forsythe, S.W. 1981. Bottomland hardwood forest communities and their relation to anaerobic soil conditions. In: Clark, J.R. and Benforado, J., eds., *Wetlands of Bottomland Hardwoods*, New York: Elsevier, pp. 187–196.

Hughes, F.M.R. 1990. Influence of flooding regimes on forest distribution and composition in the Tana River floodplain, Kenya. *Journal of Applied Ecology* 27: 475–491.

Hughes, F.M.R. 1997 Floodplain biogeomorphology. *Progress in Physical Geography* 21: 501–529.

Hughes, F.M.R., Colston, A. and Mountford, J.O. 2005. Restoring riparian ecosystems: the challenge of accommodating variability and designing restoration trajectories. *Ecology and Society* 10(1): 12.

Hupp, C.R. 1983. Geo-botanical evidence of late Quaternary mass wasting in block field areas of Virginia. *Earth Surface Processes and landforms* 8: 439–450.

Hupp, C.R. 1984. Dendrogeomorphic evidence of debris flow frequency and magnitude on Mount Shasta, California. *Environmental Geology and Water Sciences* 6: 121–128.

Hupp, C.R. 1987. Botanical evidence of floods and paleoflood history. In: Singh, V.J., ed., *Regional Flood Frequency Analysis. Proceedings of International Symposium on Flood Frequency and Risk Analysis, Baton Rouge, LA, May 1986*, Dordrecht: Reidel, pp. 355–369.

Hupp, C.R. 1988. Plant ecological aspects of flood geomorphology and paleoflood history. In: Baker, V.R., Kochel, R.C. and Patton, P.C., eds., *Flood Geomorphology*, New York: John Wiley & Sons, pp. 335–356.

Hupp, C.R. 1992. Riparian vegetation recovery patterns following stream channelization: a geomorphic perspective. *Ecology* 73: 1209–1226.

Hupp, C.R. 1999. Relations among riparian vegetation, channel incision processes and forms and large woody debris. In: Darby, S. and Simon, A., eds., *Incised River Channels*, Chichester: John Wiley & Sons, pp. 217–245.

Hupp, C.R. 2000. Hydrology, geomorphology and vegetation of Coastal Plain rivers in the southeastern United States. Special Issue. *Hydrologic Processes* 14: 2991–3010.

Hupp, C.R. and Bazemore, D.E. 1993. Spatial and temporal aspects of sediment deposition in West Tennessee forested wetlands. *Journal of Hydrology* 141: 179–196.

Hupp, C.R. and Carey, W.P. 1990. Dendrogeomorphic approach to slope retreat, Maxey Flats, Kentucky. *Geology* 18: 658–661.

Hupp, C.R. and Morris, E.E. 1990. A dendrogeomorphic approach to measurement of sedimentation in a forested wetland, Black Swamp, Arkansas. *Wetlands* 10: 107–124.

Hupp, C.R. and Osterkamp, W.R. 1985. Bottomland vegetation distribution along Passage Creek, Virginia, in relation to fluvial landforms. *Ecology* 66: 670–681.

Hupp, C.R. and Osterkamp, W.R. 1996. Riparian vegetation and fluvial geomorphic processes. *Geomorphology* 14: 277–295.

Hupp, C.R. and Osterkamp, W.R. 2013. Vegetation ecogeomorphology, dynamic equilibrium and disturbance. In: Shroder, J., Butler, D.R. and Hupp, C.R., eds., *Treatise on Geomorphology, Vol. 12, Ecologeomorphology*, San Diego, CA: Academic Press, pp. 94–106.

Hupp, C.R.and Rinaldi, M. 2007. Riparian vegetation patterns in relation to fluvial landforms and channel evolution along selected rivers of Tuscany (Central Italy). *Annals of the Association of American Geographers* 97: 12–30.

Hupp, C.R. and Simon, A. 1991. Bank accretion and the development of vegetated depositional surfaces along modified alluvial channels. *Geomorphology* 4: 111–124.

Hupp, C.R., Osterkamp, W.R. and Thornton, J.L. 1987. *Dendrogeomorphic Evidence and Dating of Recent Debris Flows on Mount Shasta, Northern California, US Geological Survey Professional Paper 1396-B*, Washington, DC: US Government Printing Office.

Hupp, C.R., Woodside, M.D. and Yanosky, T.M. 1993. Sediment and trace element trapping in a forested wetland, Chickahominy River, Virginia. *Wetlands* 13: 95–104.

Hupp, C.R., Osterkamp, W.R. and Howard, A.D., eds. 1995. *Biogeomorphology – Terrestrial and Freshwater Systems*, Amsterdam: Elsevier, 347pp.

Hupp, C.R., Pierce, A.R. and Noe, G.B. 2009. Floodplain geomorphic processes and environmental impacts of human alteration along Coastal Plain rivers, USA: *Wetlands* 29: 413–429.

Huston, M.A. 1979. A general hypothesis of species diversity. *American Naturalist* 113: 81–99.

Jacoby, G.C. and Hornbeck, J.W., eds. 1987. *Proceedings of the International Symposium on Ecological Aspects of Tree-Ring Analysis, Publication CONF 8608144*, Washington, DC: US Department of Energy, 716pp.

Johnson, W.B., Sasser, C.E. and Gosselink, J.G. 1985. Succession of vegetation in an evolving river delta, Atchafalaya Bay, Louisiana. *Journal of Ecology* 73: 973–986.

Johnson, W.C. 1994. Woodland expansion in the Platte River, Nebraska: patterns and causes. *Ecological Monographs* 64: 45–84.

Johnston, C.A. 1991. Sediment and nutrient retention by freshwater wetlands: effects on surface water quality. *Critical Reviews in Environmental Control* 21: 491–565.

Jones, H. 1956. Studies on the ecology of the river Rheidol. II. An ox-bow of the lower Rheidol. *Journal of Ecology* 44: 12–27.

Jolley, R.L., Lockaby, B.G. and Cavalcanti, G.G. 2010. Changes in riparian forest composition along a sedimentation rate gradient. *Plant Ecology* 210: 317–330.

Kalliola, R., Salo, J., Puhakka, M. and Rajasilta, M. 1991. New site formation and colonizing vegetation in primary succession on the western Amazon floodplains. *Journal of Ecology* 79: 877–901.

Karrenberg, S., Blaser, S., Kollmann, J., Speck, T. and Edwards, P.J. 2003. Root anchorage of saplings and cuttings of woody pioneer species in a riparian environment. *Functional Ecology* 17: 170–177.

Kasse, C., Bohncke, S.J.P., Vandenberghe, J. and Gabris, G. 2010. Fluvial style changes during the last glacial–interglacial transition in the middle Tisza valley (Hungary). *Proceedings of the Geologists Association*, 121(2): 180–194.

Kellerhals, R. and Church, M. 1989. The morphology of large rivers: characterization and management. In: Dodge, D.P., ed., *Proceedings of the International Large River Symposium*, Canadian Fisheries and Aquatic Science, Special Publication 106, pp. 13–30.

Kirkman, L.K. and Sharitz, R.R. 1994. Vegetation disturbance and maintenance of diversity in intermittently flooded Carolina bays in south Carolina. *Ecological Applications* 41: 177–188.

Kleiss, B.A. 1996. Sediment retention in a bottomland hardwood wetland in eastern Arkansas. *Wetlands* 16: 321–333.

Koch, J. 2009. Improving age estimates for late Holocene glacial landforms using dendrochronology – some examples from Garibaldi Provincial Park. *Quaternary Geochronology* 4(2): 130–139.

Kochel, R.C., Miller, J.R. and Ritter, D.F. 1997. Geomorphic response to minor cyclic climate changes, San Diego County, California. *Geomorphology* 19: 277–302.

Kohler, A. and Schiele, S. 1985. Veränderungen von Flora und Vegetation in den kalkreichen Fliessgewässern der Friedberger Au (bei Augsburg) von 1972 bis 1982 unter veränderten Belastungbedingungen. *Archiv für Hydrobiologie* 103: 137–199.

Kondolf, M.G., Piégay, H. and Landon, N. 2007. Changes in the riparian zone of the lower Eygues river, France, since 1830. *Landscape Ecology* 22: 367–384.

Kroes, D.E. and Hupp, C.R. 2010. The effect of channelization on floodplain sediment deposition and subsidence along the Pocomoke River, Maryland. *Journal of the American Water Resources Association* 46: 686–699.

LaMarche, V.C. Jr. 1968. *Rates of Slope Degradation as Determined from Botanical Evidence, White Mountains, California*. US Geological Survey Professional Paper 352-I, Washington, DC: US Government Printing Office, pp. 1341–1377.

Lespez, L., Clet-Pellerin, M., Davidson, R., Hermier, G., Carpentier, V. and Cador, J.-M. 2010. Middle to Late Holocene landscape changes and geoarchaeological implications in the marshes of the Dives estuary (NW France). *Quaternary International* 216: 23–40.

Lowe, B.J., Watts, R.J. and Robertson, R.A. 2010. The effect of experimental inundation and sediment deposition on the survival and growth of two herbaceous riverbank plant species. *Plant Ecology* 209: 57–69.

Maas, G.S., Macklin, M.G. and Kirkby, M.J. 1998. Late Pleistocene and Holocene River development in Mediterranean Steepland environments, Southwest Crete, Greece. In: Benito, G., Baker, V.R. and Gregory, K.J., eds., *Palaeohydrology and Environmental Change*, New York: John Wiley & Sons, pp. 153–165.

Maas, G.S., Macklin, M.G., Warburton, J., Woodward, J.C. and Meldrum, E. 2001. A 300 years history of flooding in an Andean mountain river system: the Rio Alizos, southern Bolivia. In: Maddy, D., Macklin, M.G. and Woodward, J.C., eds., *River Basin Sediment Systems: Archives of Environmental Change*, Rotterdam: Balkema, pp. 297–324.

Macklin, M., Rumsb, B. and Heap, T. 1992. Flood alluviation and entrenchment: Holocene valley-floor development and transformation in the British Uplands. *Geological Society of America Bulletin* 104: 631–643.

Magilligan, F.J. and Stamp, M.L. 1997. Historical land-cover changes and hydrogeomorphic adjustment in a small Georgia watershed, *Annals of the Association of American Geographers* 87: 614–635.

Malik, I. 2006. Contribution to understanding the historical evolution of meandering rivers using dendrochronological methods: example of the MaL a Panew River in southern Poland. *Earth Surface Processes and Landforms* 31: 1227–1245.

Marston, R.A., Girel, J., Pautou, G., Piégay, H., Bravard, J.P. and Arneson, C. 1995. Channel metamorphosis, floodplain disturbance and vegetation development: Ain river, France. *Geomorphology* 13: 121–131.

McKenney, R., Jacobson, R.B. and Wertheimer, R.C. 1995. Woody vegetation and channel morphogenesis in low gradient, gravel-bed streams in the Ozark Plateaus, Missouri and Arkansas. *Geomorphology* 13: 175–198.

Merigliano, M.F. 2005. Cottonwood understory zonation and its relation to floodplain stratigraphy. *Wetlands* 25: 356–374.

Merigliano, M.F., Friedman, J.M. and Scott, M.L. 2013. Tree-ring records of variation in flow and channel geometry. In: Shroder, J., Butler, D.R. and Hupp, C.R., eds., *Treatise on Geomorphology, Vol. 12, Ecogeomorphology*, San Diego, CA: Academic Press, pp. 145–164.

Mitsch, W.J. and Gosselink, J.G. 1993. *Wetlands*, 2nd edn., New York: Van Nostrand Reinhold.

Mizugaki, S., Nakamura, F. and Araya, T. 2006. Using dendrogeomorphology and Cs-137 and Pb-210 radiochronology to estimate recent changes in sedimentation rates in Kushiro Mire, *Northern Japan, resulting from land use change and river channelization Catena* 68: 25–40.

Müller, N. 1995. River dynamics and floodplain vegetation and their alterations due to human impact. *Archiv für Hydrobiologie, Supplement 101, Large Rivers* 9(3/4): 477–512.

Nádor, A., Sinha, R., Magyari, A., Tandon, S.K., Medzihradszky, Z., Babinszki, E., Thamó-Bozsó, E., Unger, Z. and Singh, A. 2011. Late Quaternary (Weichselian) alluvial history and neotectonic control on fl uvial landscape development in the southern Körös plain, Hungary. *Palaeogeography, Palaeoclimatology, Palaeoecology* 299: 1–14.

Naiman, R.J., Décamps, H. and Pollock, M. 1993. The role of riparian corridors in maintaining regional diversity. *Ecological Applications* 3: 209–212.

Nakamura, F. and Shin, N. 2001. The downstream effects of dams on the regeneration of riparian tree species in northern Japan. In: Dorava, J.M., Montgomery, D.R., Palcsak, B.B. and Fitzpatrick, F.A., eds., *Geomorphic Processes and Riverine Habitat*, Washington, DC: American Geophysical Union, pp. 173–181

Nakamura, F., Shin, N. and Inahara, S. 2007. Shifting mosaic in maintaining diversity of floodplain tree species in the northern temperate zone of Japan. *Forest Ecology and Management* 241: 28–38.

Nanson, G.C. and Beach, H.F. 1977. Forest succession and sedimentation on a meandering-river floodplain, northeast British Columbia, Canada. *Journal of Biogeography* 4: 229–251.

Naylor, L.A. 2005. The contributions of biogeomorphology to the emerging field of geobiology. *Palaeogeography, Palaeoclimatology, Palaeoecology* 219: 35–51.

Nilsson, C., Grelsson, G., Johansson, M. and Sperens, U. 1989. Patterns of plants species richness along riverbanks. *Ecology* 70: 77–84.

Nilsson, C., Reidy, C.A., Dynesius, M. and Revenga, C. 2005. Fragmentation and flow regulation of the worlds large river systems: *Science* 308: 405–408.

Noe, G.B. and Hupp, C.R. 2009. Retention of riverine sediment and nutrient loads by Coastal Plain floodplains. *Ecosystems* 12: 728–746.

Odum, E.P. 1971. *Fundamentals of Ecology*, 3rd edn., Philadelphia, PA: W.B. Saunders, 574pp.

Ollero, A. 2010. Channel changes and floodplain management in the meandering middle Ebro River, Spain. *Geomorphology* 117: 247–260.

Olson, J.S. 1958. Lake Michigan dune development 2. Plants as agents and tools in geomorphology. *Journal of Geology* 66(4): 345–351.

Osterkamp, W.R. and Hupp, C.R. 1984. Geomorphic and vegetative characteristics along three Northern Virginian streams. *Geological Society of America Bulletin* 95: 1093–1101.

Osterkamp, W.R. and Hupp, C.R. 2010. Fluvial processes and vegetation – glimpses of the past, present and future: *Geomorphology* 116: 274–285.

Osterkamp, W.R., Hupp, C.R. and Stoffel, M. 2012 The interactions between vegetation and erosion: new directions for research at the interface of ecology and geomorphology. *Earth Surface Processes and Landforms* 37: 23–36.

Patterson, G.G., Speiran, G.K. and Whetstone, B.H. 1985. *Hydrology and Its Effects on Distribution of Vegetation in Congaree National*

Monument, South Carolina, US Geological Survey Water-Resources Investigations Report 85-4256, Washington, DC: US Government Printing Office, 40pp.

Pautou, G. 1984. L'organisation des forêts alluviales dans l'axe rhodanien entre Genève et Lyon; comparaison avec d'autres systèmes fluviaux. *Documents de Cartographie Ecologique* 27: 43–64.

Pautou, G. and Arens, M.-F. 1994. Theoretical habitat templets, species traits and species richness: floodplain vegetation in the Upper Rhône River. *Freshwater Biology* 31: 507–522.

Peiry, J.-L., Salvador, P.-G. and Nouguier, F. 1994. L'incision des rivières dans les Alpes du Nord: état de la question. *Revue de Géographie de Lyon* 69: 47–56.

Petts, G.E. and Bravard, J.P. 1996. A drainage basin perspective. In: Petts, G.E. and Amoros, C., eds., *Fluvial Hydrosystems*, London: Chapman & Hall, pp. 13–35.

Phillips, J.D. 2009. Biological energy in landscape evolution. *American Journal of Science* 309: 271–289.

Phipps, R.L. 1983. Streamflow of the Occoquan River in Virginia as reconstructed from tree-ring series. *Water Resources Bulletin* 19: 735–743.

Pickett, S.T.A. 1980. Non-equilibrium coexistence of plants. *Bulletin of the Torrey Botanical Club* 107: 238–248.

Piégay, H. and Gurnell, A.M. 1997. Large woody debris and river geomorphological pattern: examples from S.E. France and S. England. *Geomorphology* 19: 99–116.

Piégay, H., Walling, D.E., Landon, N., He, Q., Liébault, F. and Petiot, R. 2004. Contemporary changes in sediment yield in an alpine mountain basin due to afforestation (the upper Drôme in France). *Catena* 55: 183–212.

Piégay, H., Hupp, C.R., Citterio, A., Dufour, S., Moulin, B. and Walling, D.E. 2008. Spatial and temporal variability in sedimentation rates associated with cut-off channel infill deposits: Ain River, France. *Water Resources Research* 44(5): W05420.

Pierson, T.C. 2007. Dating young geomorphic surfaces using age of colonizing Douglas fir in southwestern Washington and northwestern Oregon, USA. *Earth Surface Processes and Landforms* 32: 811–831.

Pike, A.S. and Scatena, F.N. 2010. Riparian indicators of flow frequency in a tropical montane stream network. *Journal of Hydrology* 382: 72–87.

Provansal, M., Villiet, J., Eyrolle, F., Raccasi, G., Gurriaran, R. and Antonelli, C. 2010. High-resolution evaluation of recent bank accretion rate of the managed Rhone: a case study by multi-proxy approach. *Geomorphology* 117: 287–297.

Resh, V.H., Brown, A.V., Covich, A.P., Sheldon, A.L., Wallace, J.B. and Wissman, R.C. 1988. The role of disturbance in stream hydrology. *Journal of the North American Benthological Society* 7: 433–455.

Ribicki, N.B., and Carter, V. 1986. Effect of sediment depth and sediment type on the survival of Vallisneria americana Michx grown from tubers. *Aquatic Botany* 24: 233–240.

Rinaldi, M. 2003. Recent channel adjustments in alluvial rivers of Tuscany, Central Italy. *Earth Surface Processes and Landforms* 28: 587–608.

Roberts, J. and Ludwig, J.A. 1991. Riparian vegetation along current-exposure gradients in floodplain wetlands of the River Murray, Australia. *Journal of Ecology* 79: 117–127.

Rojo, L.D., Mehl, A.E., Paez, M.M., and Zárate, M.A. 2012. Mid- to Late Holocene pollen and alluvial record of the arid Andean piedmont between 33° and 34°S, Mendoza, Argentina: inferences about floodplain evolution. *Journal of Arid Environments* 77: 110–122.

Rollet, A.J., Piégay, H., Bornette, G., Dufour, S. and Persat, H. 2014. Assessment of consequences of sediment deficit on a gravel river-bed downstream of dams in restoration perspectives: application of a multicriteria, hierarchical and spatially explicit diagnosis. *River Research and Applications* 30: 939–953.

Rosales-Godoy, J.R., Petts, G. and Salo, J. 1999. Riparian flooded forests of the Orinoco and Amazon basins: a compariative review. *Biodiversity and Conservation* 8: 551–586.

Ross, K.M., Hupp, C.R. and Howard, A.D. 2004. Sedimentation in floodplains of selected tributaries of the Chesapeake Bay. In: Bennett, S.J and Simon, A., eds., *Riparian Vegetation and Fluvial Geomorphology, Water Science and Applications 8*, Washington, DC: American Geophysical Union, pp. 187–208.

Rubtsov, M.V. and Salmina, I.U.N. 1983. Use of buried Salix thickets for determination of the chronology of floodplain deposits. *Lesovedinie* 69–70.

Sambaré, O., Bognounou, F., Wittig, R. and Thiombiano, A. 2011. Woody species composition, diversity and structure of riparian forests of four watercourses types in Burkina Faso. *Journal of Forestry Research* 22: 145–158.

Schumm, S.A. 1985. Patterns of alluvial rivers. *Annual Review of Earth and Planetary Sciences* 13: 5–27.

Schweingruber, F.H. 1988. *Tree Rings: Basics and Applications of Dendrochronology*, Dordrecht: Kluwer.

Scott, M.L., Friedman, J.M. and Auble, G.T. 1996. Fluvial processes and the establishment of bottomland trees. *Geomorphology* 14: 327–339.

Scott, M.L., Auble, G.T. and Friedman, J.M. 1997. Flood dependency of cottonwood establishment along the Missouri River, Montana, USA. *Ecological Applications* 7: 677–690.

Sharitz, R.R. and Mitsch, W.J. 1993. Southern floodplain forests. In: Martin, W.G., Boyce, S.G. and Echternacht, A.C., eds., *Biodiversity of the Southeastern United States, Lowland Terrestrial Communities*, New York: John Wiley & Sons, pp. 311–372.

Shroder, J.F. 1978. Dendrogeomorphological analysis of mass movement of Table Cliffs Plateau, Utah. *Quaternary Research* 9: 168–185.

Sigafoos, R.S. 1964. *Botanical Evidence of Floods and Flood-plain Deposition, US Geological Survey Professional Paper 485-A*, Washington, DC: US Government Printing Office.

Sigafoos, R.S., and Hendricks, E.L. 1961. *Botanical Evidence of the Modern History of Nisqually Glacier, Washington, US Geological Survey Professional Paper 387-A*, Washington, DC: US Government Printing Office.

Simon, A. and Hupp, C.R. 1987. Geomorphic and vegetative recovery processes along modified Tennessee streams: an interdisciplinary approach to disturbed fluvial systems. In: *Forest Hydrology and Watershed Management, IAHS-IASH Publication No. 167*, Washington, DC: International Association of Hydrological Sciences, pp. 251–262.

Simon, A. and Hupp, C.R. 1992. *Geomorphic and Vegetative Recovery Processes Along Modified Stream Channels of West Tennessee, US Geological Survey Open File Report 91-502*, Nashville, TN: US Geological Survey.

Singer, M.B., Johnstone, L., Dufour, S., Stella, J.C., Piégay, H. and Wilson R.J.S. 2012. Water uptake and growth responses to drought in co-occurring riparian tree species. *Ecohydrology* 6: 402–412.

Sousa, W.P. 1984. The role of disturbance in natural communities. *Annual Review of Ecology and Systematics* 15: 353–391.

Sparks, R.E., Bayley, P.B., Kohler, S.L. and Osborne, L.L. 1990. Disturbance and recovery of large floodplain rivers. *Environmental Management* 14: 699–709.

St. George, S. and Nielsen, E. 2003. Paleoflood records for the Red River, Manitoba, Canada, derived from anatomical tree-ring signatures. *The Holocene* 13: 547–555.

Steiger, J., Gurnell, A.M., Ergenzinger, P. and Snelder, D. 2001. Sedimentation in the riparian zone of an incising river. *Earth Surface Processes and Landforms* 26: 91–108.

Steiger, J., Tabacchi, E., Dufour, S., Corenblit, D. and Peiry, J.L. 2005. Hydrogeomorphic processes affecting riparian habitat within alluvial channel-floodplain river systems: a review for the temperate zone. *River Research and Applications* 21: 719–737.

Stella, J.C., Hayden, M.K., Battles, J.J., Piégay, H., Dufour, S. and Fremier, A.K. 2011. The role of abandoned channels as refugia for sustaining pioneer riparian forest ecosystems. *Ecosystems* 14: 776–790.

Stella, J.C., Riddle, J., Piégay, H., Gagnage, M. and Trémélo, M.-L. 2013. Climate and local geomorphic interactions drive patterns of riparian forest decline along a Mediterranean Basin river. *Geomorphology* 202: 101–114.

Stoffel, M. and Bollschweiler, M. 2008. Tree-ring analysis in natural hazards research – an overview. *Natural Hazards and Earth System Sciences* 8: 187–202.

Stoffel, M., Lievre, I., Conus, D., Grichting, M.A., Raetzo, H., Gärtner, H.W. and Monbaron, M. 2005. 400 yeaars of debris flow activity and triggering weather conditions: Ritigarben, Valais, Switzerland. *Arctic, Antarctic and Alpine Research* 37: 387–395.

Stoffel, M., Conus, D., Grichting, M.A., Lievre, I. and Maitre, G. 2008 Unraveling the patterns of late Holocene debris-flow activity on a cone in the Swiss Alps: chronology, environment and implications for the future. *Global and Planetary Change* 60: 222–234.

Stoffel, M., Casteller, A., Luckman, B. and Villalba, R. 2012. Spatiotemporal analysis of channel wall erosion in ephemeral torrents using tree roots – and example from the Patagonian Andes. *Geology* 40: 247–250.

Stotts, S., O'Neal, M., Pizzuto, J., and Hupp, C. 2013. Exposed tree root analysis as a dendrogeomorphic approach to estimating bank retreat at the South River, Virginia. *Geomorphology*, 223: 10–18.

Strunk, H. 1997. Dating of geomorphological processes using dendrogeomorphological methods. *Catena* 31: 137–151.

Swanson, F.J., Kratz, T.K., Caine, N. and Woodmansee, R.G. 1988. Landform effects on ecosystem patterns and processes. *BioScience* 38(2): 92–98.

Tabacchi, E. and Planty-Tabacchi, A.M. 2005. Exotic and native plant community distributions within complex riparian landscapes: a positive correlation. *Ecoscience* 12: 423–434.

Tabacchi, E., Correll, D.L., Hauer, R., Pinay, G., Planty-Tabacchi, A.M. and Wissmar, R.C. 1998. Development, maintenance and role of riparian vegetation in the river landscape. *Freshwater Biology* 40: 497–516.

Thornes, J.B., ed. 1990. *Vegetation and Erosion: Processes and Environments, British Geomophological Research Group Symposia Series*, Chichester: John Wiley & Sons, 518pp.

Turner, F., Tolksdorf, J.F., Viehberg, F., Schwalb, A., Kaiser, K., Bittmann, F., von Bramann, U., Pott, R., Staesche, U., Breest, K. and Veil, S. 2013. Late glacial/early Holocene fluvial reactions of the Jeetzel river (Elbe valley, northern Germany) to abrupt climatic and environmental changes. *Quaternary Science Reviews* 60: 91–109.

Uunila, L.S. 1997. Effects of river ice on bank morphology and riparian vegetation along Peace River, Clayhurst to Fort Vermillon. In: *Proceedings of the 9th Workshop on the Hydraulics of Ice Covered Rivers*, September 24–26, Fredericton, NB, pp. 315–334. http://cripe.civil.ualberta.ca/proceedings/cripe-workshop09.html (accessed 12 October 2015).

van der Valk, A.G. and Bliss, L.C. 1971. Hydrarch succession and net primary production of oxbow lakes in central Alberta. *Canadian Journal of Botany* 49: 1177–1199.

Van Looy, K., Meire, P. and Wasson, J.G. 2008. Including riparian vegetation in the definition of morphologic reference conditions for large rivers: a case study for Europe's Western plain. *Environmental Management* 41: 625–639.

Vannote, R.L., Minshall, G.W., Cummins, K.W., Sedell, J.R. and Cushing, C.E. 1980. The river continuum concept. *Canadian Journal of Fisheries and Aquatic Sciences* 37: 130–137.

Vartapetian, B.B. and Jackson, M.B. 1997. Plant adaptations to anaerobic stress. *Annals of Botany* 79(Suppl. A): 3–20.

Viles, H.A. 1995. Ecological perspectives on rock surface weathering: towards a conceptual model. *Geomorphology* 13: 21–35.

Viles, H.A., ed. 1988. *Biogeomorphology*, Oxford: Blackwell.

Viles, H.A., Naylor, L.A., Carter, N.E.A. and Chaput, D. 2008. Biogeomorphological disturbance regimes: progress in linking ecological and geomorphological systems. *Earth Surface Processes and Landforms* 33: 1419–1435.

Ward, J.V., Tockner, K., Edwards, P.J., Kollman, J., Bretschko, G., Gurnell, A.M., Petts, G.E. and Rossaro, B. 1999. A reference river system for the Alps: the 'Fiume Tagliamento'. *Regulated Rivers – Research and Management* 15: 63–75.

Ward, J.V., Tockner, K., Uehlinger, U. and Malard, F. 2001. Understanding natural patterns and processes in river corridors as the basis for effective river restoration. *Regulated Rivers – Research and Management* 17: 311–323.

Wharton, C.H., Kitchens, W.M., Pendleton, E.C. and Sipe, T.W. 1982. *The Ecology of Bottomland Hardwood Swamps of the Southeast: a Community Profile, US Fish and Wildlife Service FWS/OBS-81/37*, Washington, DC: Biological Services Program.

Wheaton, J.M., Gibbins, C., Wainwright, J., Larsen, L. and McElroy, B., eds. 2011. Multiscale Feedbacks in Ecogeomorphology. *Special Issue, Geomorphology* 126.

White, P.S. and and Pickett, S.T.A. 1985. Natural disturbance and patch dynamics: an introduction. In: Pickett, S.T.A. and White, P.S., eds., *The Ecology of Natural Disturbance and Patch Dynamics*, San Diego, CA: Academic Press, pp. 3–13.

Willard, D.A. Bernhardt, C.E., Korejwo, D.A. and Myers, S.R. 2005. Impact of millennial-scale Holocene climate variability on eastern North American terrestrial ecosystems: pollen-based climatic resconstruction. *Global and Planetary Change* 47: 17–35.

Willard, D.A., Bernhardt, C.E., Brown, R., Landacre, B. and Townsend, P. 2011. Development and application of a pollen-based paleohydrologic reconstruction from the Lower Roanoke River Basin, North Carolina, USA. *The Holocene* 21: 305–317

Williams, G.P. and Wolman, M.G. 1984. *Downstream Effects of Dams on Alluvial Rivers, US Geological Survey Professional Paper 1286*, Washington, DC: US Government Printing Office.

Willms, J., Rood, SB., Willms, W. and Tyree, M. 1998. Branch growth of riparian cottonwoods: a hydrologically sensitive dendrochronological tool. *Trees* 12: 215–223.

Wohl, E. and Merritt, DM. 2008. Reach-scale channel geometry of mountain streams. *Geomorphology* 93: 168–185.

Wyant, J.G. and Ellis, J.E. 1990. Compositional patterns of riparian woodlands in the Rift Valley of northern Kenya. *Vegetatio* 89: 23–37.

Wyzga, B. 1999. Estimating mean flow velocity in channel and floodplain areas and its use for explaining the pattern of overbank deposition and floodplain retention. *Geomorphology* 28: 281–297.

Yanosky, T.M. 1982. Hydrological inferences from ring widths of flood-damaged trees, Potomac River, Maryland. *Environmental Geology and Water Sciences* 4: 43–52.

Yanosky, T.M. 1983. *Evidence of Floods on the Potomac River from Anatomical Abnormalities in the Wood of Floodplain Trees, US Geological Survey Professional Paper 1296*, Washington, DC: US Government Printing Office.

Yanosky, T.M., Hupp, C.R. and Hackney, C.T. 1995. Chloride concentrations in growth rings of Taxodium distichum in a saltwater-intruded estuary. *Ecological Applications* 5: 785–792.

Zimmermann, R.C. and Thom, B.G. 1982. Physiographic plant geography. *Progress in Physical Geography* 6: 45–59.

SECTION V

Analysis of Processes and Forms: Water and Sediment Interactions

CHAPTER 11

Channel form and adjustment: characterization, measurement, interpretation and analysis

Andrew Simon[1], Janine Castro[2] and Massimo Rinaldi[3]
[1]Cardno ENTRIX, Oxford, MS, USA
[2]US Fish and Wildlife Service, Portland, OR, USA
[3]Università degli Studi di Firenze, Florence, Italy

11.1 Introduction

Channel form, or morphology, has long been recognized as a diagnostic tool in evaluating fluvial landforms. Since Davis (1909) conceptualized the temporal aspect of channel and drainage basin evolution in the 'cycle of erosion', geographers, geologists and geomorphologists have used channel form as a parameter in the classification, analysis and prediction of fluvial response. Davis's view of fluvial landscapes was simplistic but, in combination with the detailed measurements of channel forms and processes in the studies by Gilbert (1914), we can envision these works as representing opposite approaches by which to direct future work. Davis's work represents large-scale, qualitative assessments of channel form by which inferences about smaller scale processes were advanced. Conversely, Gilbert's work represents the use of quantitative measurements by which inferences about larger scale processes were advanced. The implied links between channel form and process have been central in understanding fluvial geomorphology and, as such, have been the topic of many textbooks (e.g. Leopold *et al.* 1964; Morisawa 1968; Gregory and Walling 1973; Schumm 1977; Richards 1982; Knighton 1998).

Channel form includes aspects of the shape of a stream in profile, cross-section and planform. Profile characteristics include channel-bed gradient and valley slope and features such as pools, riffles and cascades. Cross-sectional characteristics include channel width and depth and features such as the bed, bars, banks, floodplains and terraces. Planform characteristics include sinuosity, meander wavelength and belt width and features such as meanders, braids and abandoned channels. Channel form measurements and descriptions can be used in combination with other attributes of a stream system, such as riparian vegetation and character of the boundary sediments, to infer dominant trends in channel processes and response (Simon and Hupp 1986; Simon 1989a; Montgomery and Buffington 1997; Elliott *et al.* 1999). However, using gross channel form in isolation to quantitatively predict channel behaviour, such as channel adjustments, system disturbances or rates of sediment transport, without rigorous analysis of channel processes is unsound (Miller and Ritter 1996). Hence the key to using channel form in the analysis of fluvial landforms must be based on either (i) measurements of parameters that aid in quantifying channel processes, such as flow hydraulics, sediment transport and bank stability, or (ii) observations of diagnostic characteristics that provide information on active channel processes. In turn, measurements should either directly or indirectly lead to analysis of those forces acting on the channel boundary and those forces resisting entrainment of boundary sediments. Change in channel form is a matter of the applied forces overcoming resistance.

The purpose of this chapter is to provide a synthetic overview of available techniques, methods and parameters for characterizing and measuring channel forms and analysing and interpreting changes over time. Hence the chapter is organized into two major components, characterization and measurement, followed by analysis and interpretation. Synthetic review of analysis of channel changes, including measurement of some of the parameters that aid in quantifying channel processes responsible for morphological changes, is also included.

Because of the broad scope of this chapter, it is not possible to address specifically all channel form measurement and analysis techniques; however, there are several other chapters that focus on specific components, including: Chapter 4, Using historical data in fluvial geomorphology; Chapter 6, Analysis of aerial photography and other remotely sensed data for fluvial geomorphology and river science; Chapter 13, Measuring bed sediment; and Chapter 15, Sediment transport.

11.2 Characterization and measurement

Characterization and measurement of channel form provide information for river classification and studies on past and future channel changes. Following the basic tripartive division

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

of channel patterns provided by Leopold and Wolman (1957) of braided, meandering and straight, various morphological classifications have been developed (e.g. Schumm 1977, 1981; Church 1992; Thorne 1997; Montgomery and Buffington 1997; Fuller *et al.* 2013; see also Chapter 7). For the scope of this chapter, an initial distinction is drawn between bedrock and alluvial channels. Channels formed in bedrock have a discontinuous or thin alluvial cover and cannot substantially widen, lower or shift their bed without eroding bedrock (Turowski *et al.* 2008), whereas channels formed in sediment that can be eroded, transported and deposited by the flow can be classified as alluvial or 'self-formed' (Thorne 1997). Alluvial channels adjust their form to the driving variables (i.e. water and sediment flow) and their interactions with the boundary characteristics, whereas forms and dimensions of bedrock channels are controlled by geological and structural factors.

Alluvial rivers display a wide spectrum of channel forms and morphological units on a variety of landforms, such as alluvial fans, confined alluvial valleys and wide alluvial valleys, that can be identified and classified during stream reconnaissance. They include bed morphology (i.e. cascade, step-pool, plane bed, riffle-pool, dune ripples) (Montgomery and Buffington 1997; Halwas and Church 2002), mid-channel or bank-attached sedimentary features (such as various types of bars, islands, benches, berms) (Brierley and Fryirs 2005) and fluvial forms in the alluvial plain (floodplain, terraces, secondary or abandoned channels, meander scars, scroll bars, oxbow lakes, braided deposits).

The channel, or 'active channel', includes the frequently submerged portion of the bed and all mid-channel or bank attached sedimentary features, whereas floodplains are inundated during larger, less frequent events. Identification of, and distinction among, the channel, floodplain and terraces during field surveys is of particular importance because it directly relates to disturbance regimes and rates of morphological change. Terraces represent abandoned floodplains and are found at various elevations, which are generally controlled by either the amount of channel incision that has subsequently occurred or by geological events such as uplift. Recently formed terraces are common and are typically the result of channel incision due to land management practices, channelization projects or other human disturbances (e.g. sediment mining, dams).

Streams affected by bed degradation are described as 'incised river channels'. They are ubiquitous features of landscapes disturbed by environmental changes and exhibit a series of typical channel features and processes, such as recent terraces and general lack of active floodplains, severe bank erosion and widening with unstable banks dominated by mass failures and deficit of sedimentary features (Simon 1989a; Darby and Simon 1999; Simon and Rinaldi 2006).

As mentioned above, correctly identifying characteristic forms and their underlying fluvial processes is necessary to characterize and classify river channels accurately. An initial reconnaissance-level assessment can be rapidly carried out at a large number of sites through a qualitative evaluation procedure where diagnostic criteria are used. A stream reconnaissance survey consists of a systematic collection of information on morphological features using a pro forma checklist of quantitative measurements and qualitative observations (Thorne *et al.* 1996). Many schemes for evaluating river channel morphology based on a single reconnaissance survey are available (e.g. Simon and Downs 1995; Thorne *et al.* 1996; Downs and Thorne 1996; Thorne 1998; Rinaldi 2008).

In addition to field survey, a second type of tool for characterization of channel forms is represented by aerial photo-interpretation and geomorphological mapping. Satellite images, structural light detection and ranging (LiDAR) or a specific low-altitude flight can help to integrate field observations and allow a broad, catchment-scale, rapid assessment of river channel characteristics (see also Chapter 6) Then, a more accurate stream survey can be performed on representative or selected critical sites by using more detailed approaches (Thorne 1998). Aerial photo-interpretation allows for an effective identification of channel features and fluvial forms in the alluvial plain. This tool is more suitable for relatively large, alluvial rivers, given the size limitation and errors in interpreting aerial photographs. Mapping is a widely used methodology in geomorphological studies and is broadly used to represent fluvial forms in the alluvial plain. However, mapping of dynamic in-channel morphological units and sedimentary features is merely a static representation of an instantaneous situation, hence studies of channel changes require multiple years of photograph and map interpretation that represent decades worth of change (see the next section). Digital mapping with GIS software allows for organized and systematic representation, updating and measurements of changes. Furthermore, recent advances in remote sensing technologies allow for mapping floodplain topography and vegetation cover by the simultaneous use of both LiDAR and compact airborne spectral imager (CASI) remote sensing data (Geerling *et al*. 2009).

Whereas characterization and measurement of channel forms described in this section can apply to both bedrock and alluvial channels, analysis and interpretation of channel changes discussed in the subsequent section will focus on alluvial channels, as their form can change more substantially over time and they are of greater societal interest on engineering time-scales.

Longitudinal profile and bed elevation characterization and measurement

The term 'longitudinal profile' refers to a graphical 2D representation of bed morphology, where bed elevation is plotted against longitudinal distances downstream along the channel. Bed elevation can refer to the deepest point in the channel (thalweg) or, alternatively, to the mean bed elevation. In the latter case, mean bed elevation is obtained by the measurement of a series of cross-sections along a given reach. The objective of measuring bed elevation is to gain indirect information on stream energy by the channel slope and to determine the inundation relationship between the channel, floodplain and terraces.

Longitudinal surveys provide the data needed for a number of uses, including relative bed elevation, thalweg slope, bankfull slope, valley slope and habitat units. Derived attributes from the longitudinal profile include the pattern of undulations in the bed profile (e.g. pool and riffle or pool and step spacing, residual pool depth) and breaks in slope. Of the longitudinal profile parameters, slope is the most predominately used for hydraulic models and morphological classification. Slope is used to calculate channel velocity and discharge at various stages, stream power, shear stress and other parameters that begin to quantify channel processes. Generally, a longitudinal survey will start at a stable point within the stream channel. For geomorphic investigations, long profiles will usually include (at a minimum) points on the same, repeated geomorphic features going downstream (such as the heads of riffles) and important slope breaks. For input to hydraulic models such as HEC-RAS, the long profile should include, and cross-sections should be placed at, all hydraulic controls.

Longitudinal profile and bed elevation characterization

The type of geomorphic surfaces that are encountered during survey should be noted. Sufficient detail should be provided to define clearly bed such features as steps, pools, riffles, glides and any unique conditions. Surveys of a thalweg profile should ideally encompass a reach at least 6–30 channel widths in length, especially if the purpose is to determine channel gradient, although this is not always practical. Surveys of this length will generally include a series of pool-riffle or step-pool sequences. At least two complete sequences should be included with the final calculation of slope, comprising a linear regression between distance and elevation. Generally, lower stream slopes require more extensive surveys. Site conditions may limit the extent of the survey or unique conditions may require a more expansive survey.

A total station or laser level is preferred for longitudinal profiles because of the great distances that can be obtained for individual shots and because of the sensitivity of channel slope in various hydraulic models. Using the appropriate equipment is especially critical in areas that are flat. In steep channels (>5%), a hand level that provides a derived accuracy of 0.25% for channel gradient may be sufficient for a reconnaissance-level survey; however, this is unacceptable for lower gradient stream channels, especially for those less than 2%. More specific guidelines on survey techniques can be found in Harrelson *et al.* (1994). In recent years, GPS has largely replaced traditional survey techniques, allowing for rapid and precise surveys and obtaining directionally absolute coordinates.

Where flows are too deep to obtain measurements of bed elevation, survey shots of the water surface (edge of water) or bathymetric surveys by sonar systems or echo sounding can be used.

Measurement of longitudinal profile and bed elevation change

Studies regarding changes in longitudinal profile, and specifically bed elevation, usually start from the acquisition of existing longitudinal profiles of bed elevation or topographic surveys of cross-sections. In some relatively large rivers, extensive historical series of topographic surveys can be available from past river training projects or for flood monitoring. Superimposition of bed profiles or minimum bed elevation extracted from cross-sections can be used to identify the type and amount of changes (e.g. Agnelli *et al.* 1998; Rinaldi and Simon 1998).

Abundant data are available at stream gauging stations, where a specific gauge analysis can be carried out. This analysis consists in the identification of water level changes through time for a range of specified constant discharges, indicating possible bed-level changes and/or changes in bedforms, bars and flow resistance. Mean channel bed elevations can be obtained by subtracting mean flow depth from water surface elevation for each discharge measurement where flows are at bankfull stage or below. This method is described in detail in Chapter 4 and by Jacobson (1995). If this procedure is used for at least several years of record, it should be possible to determine if bed-level adjustment is ongoing. A similar method is employed at gauging stations by Wilson and Turnipseed (1994) to obtain minimum bed elevations. Maximum flow depth of the annual minimum daily stage is subtracted from the water-surface elevation of that stage, resulting in minimum thalweg elevations with time. Alternatively, minimum annual river stage at a gauging station can provide information on trends in bed elevation.

In the absence of historical data, field evidence collected during stream reconnaissance can provide a gross estimation of the order of magnitude of changes in bed elevation. Field evidence can include: (i) exposure of bridge piles or other static structures; (ii) differences in elevation between homologous geomorphic surfaces; (iii) in-channel deposition of fine sediment or lack of in-channel deposition; (iv) dying riparian vegetation or root zones well above the water surface; (v) erosion of both stream banks; and (vi) groundwater drainage into the channel through the banks. For example, the difference in level between a low terrace identified as being a floodplain from past aerial photographs or other sources and the present active floodplain can be used to gain an approximate evaluation of the amount of incision (Rinaldi 2003). Similarly, the difference in elevation between vegetated terraces deriving from abandoned channel surfaces and the top of active channel bars (Liébault *et al.* 2012) or the top of a gravel lens visible on an eroding bank and the top of gravel in the present bed can be taken as an estimation of incision. Monitoring of present and future changes in bed elevation can be achieved by periodic measurements through one of the above-mentioned methods.

Temporal changes in bed elevation can be represented in two ways: (i) multi-temporal longitudinal profiles and (ii) bed-level adjustments at-a-site.

Multi-temporal longitudinal profiles (Fig. 11.1) provide direct information on spatio-temporal distribution of changes and capture temporal changes in bed slope (e.g. Rinaldi and Simon 1998). In the case of historical profiles or cross-sections, various sources of error need to be considered, including accuracy of old surveys, the elevation datum for each profile and identification of common reference points. When longitudinal profiles are compared, bed elevation data reported in historical profiles should be carefully evaluated. Comparison of mean bed elevation is usually more representative of the overall bed changes, whereas changes in thalweg elevation are more influenced by local conditions. Furthermore, the distances along the river channel may change due to variations of the planimetric position. In this case, some common points need to be identified along the longitudinal profiles (e.g. bridges or other fixed structures) and distances among them need to be corrected.

Plotting the changes in bed elevation for different time intervals versus the distances downstream also provides an effective way to visualize the amount and spatial distribution of bed elevation changes (Rinaldi and Simon 1998). Bed-level adjustments at a site, obtained by plotting bed elevation (or minimum annual river stage) versus time, provide detailed information on the temporal trend or trajectory of change in a single site of the fluvial system and allow identification of phases of adjustment (Fig. 11.2). They can be reconstructed by various sources of data, including extraction of bed elevation at a given site from multi-temporal longitudinal profiles, cross-sections or specific gauge analysis. As for longitudinal profiles, bed elevation can

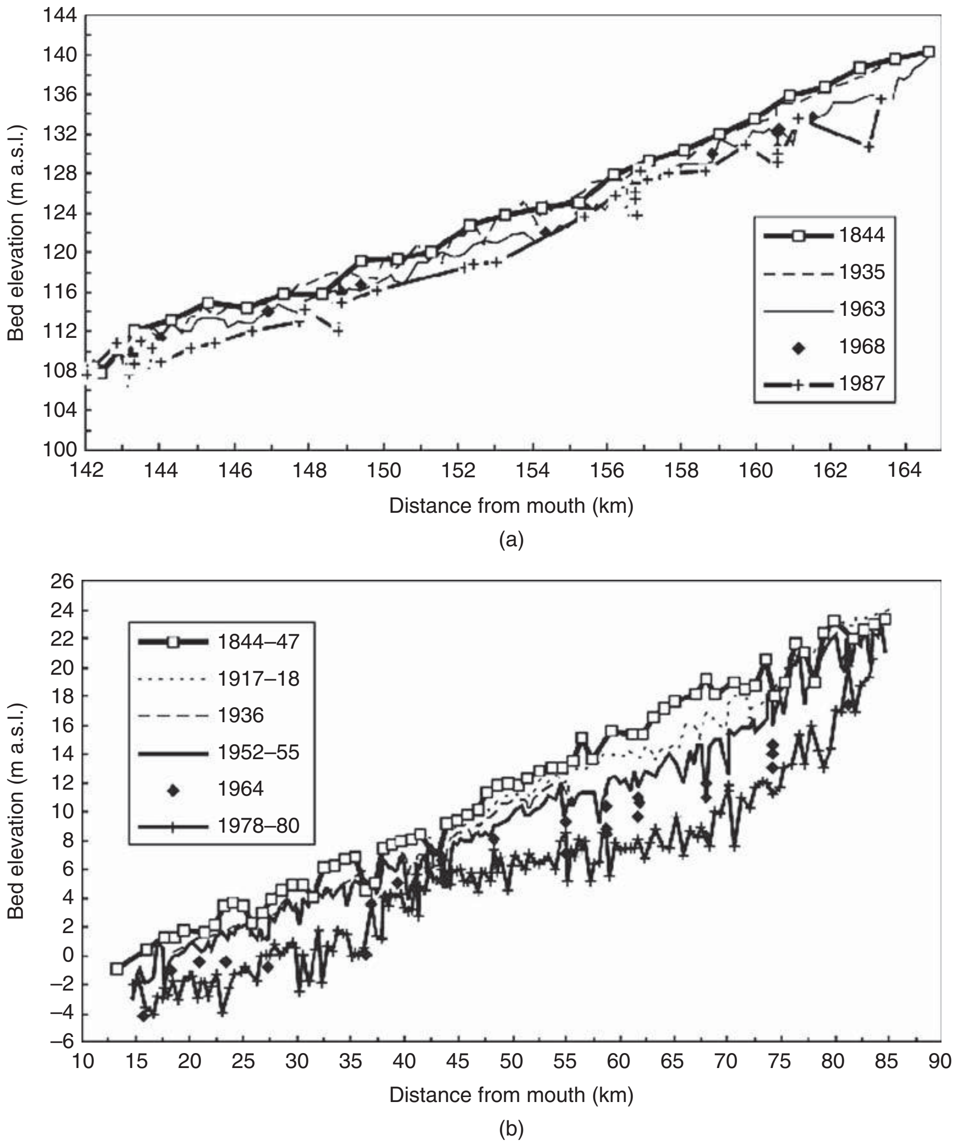


Figure 11.1 Multi-temporal longitudinal profiles: examples from the Arno River (Central Italy). Rinaldi and Simon, 1998. Reproduced with permission of Elsevier.

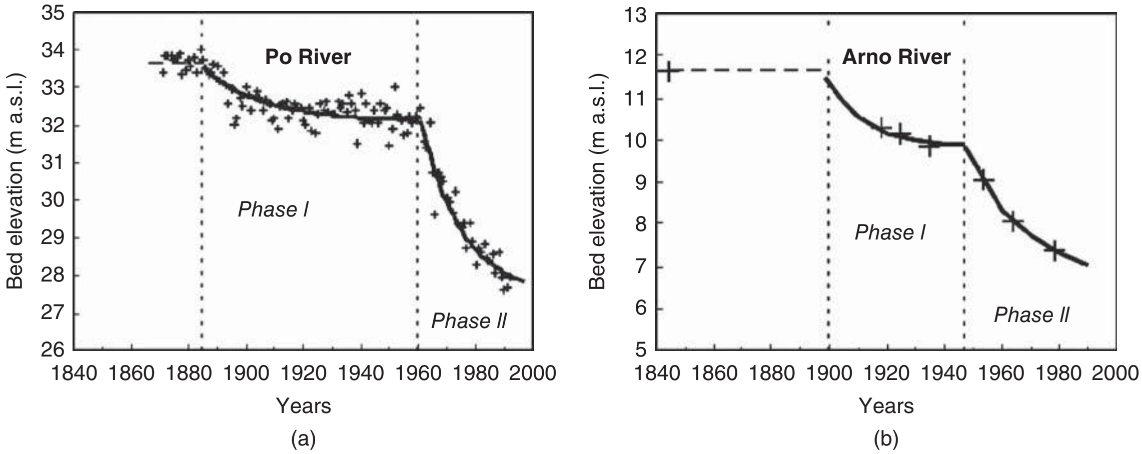


Figure 11.2 Trends of bed-level adjustments at a site and identification of phases of bed changes: examples from (a) Po River (minimum annual river stage at the gauging station of Cremona) and (b) Arno River (minimum bed elevation extracted from cross-sections of different years). Surian and Rinaldi, 2003. Reproduced with permission of Elsevier.

refer to the minimum elevation (thalweg) or to the mean bed elevation. Interpretation of temporal trends can be supported by using various mathematical functions (see the next section).

Cross-sectional characterization and measurement

The term 'cross-section' refers to a graphical 2D representation of channel morphology that is perpendicular, rather than parallel, to the flow direction along which distances and elevations are surveyed and then plotted. Cross-sectional surveys provide the data needed for a number of uses, including channel width and depth, wetted perimeter, bank height and angle and the presence, elevation and extent of floodplain and adjacent terraces. Derived attributes for the channel include cross-sectional area, average depth, hydraulic radius and width-to-depth ratio. The objectives of measuring cross-sections are to provide basic information for morphological characterization and classification and input data for hydraulic geometry, bank stability and hydrodynamic modelling. Furthermore, these values can be combined with longitudinal and hydraulic data to calculate channel velocity and discharge at various stages, stream power, shear stress and other parameters that are useful to quantify channel processes.

To calculate cross-sectional parameters, it is necessary to refer to a given flow stage (the bankfull stage is often used as the reference elevation) for calculation of hydraulic geometry. Once cross-sections have been measured, maximum depth or mean depth can be calculated. Maximum depth is obtained as the difference between flow stage and minimum bed elevation (thalweg), whereas mean depth is obtained as the difference between the selected flow stage and mean bed elevation or equivalently by the ratio between cross-sectional area and width. Mean bed elevation can be obtained as the average elevation of the points referred to the channel bed, starting from the bank toe (banks are generally excluded from this calculation).

While hydraulic geometry parameters generally refer to the bankfull level, in the case of incised channels with one or more levels of terraces, the overall cross-section can be measured to characterize the morphology of the whole channel cross-section (e.g. Elliott *et al*. 1999; Rinaldi 2003).

Cross-sectional characterization

Techniques employed for cross-section measurements are very similar to those for longitudinal profile and bed elevation and include topographic surveys and GPS or sonar systems for deep rivers. Generally, a cross-sectional survey should start on the floodplain/low terrace interface (in non-incised streams) or higher terrace surface and proceed across the floodplain, down the bank and across the channel, finishing on the opposite side of the valley. Sufficient detail should be provided to define clearly floodplain topography, natural levee deposits, bank form such as the vertical face and any failed debris, the bank-toe, the edge of water, thalweg and any bar surfaces. Notes should also be made of the existence, type and abundance of riparian vegetation on floodplain, bank and bar surfaces. Transitions between upland and riparian species should be clearly indicated, as these mark the zone of relative inundation frequency.

Depending on the level of accuracy desired and the site conditions, total stations, GPS, laser levels, transits, hand levels or level lines can be used to survey. Laser levels with remote sensors allow a single surveyor to collect cross-sectional data or several surveyors to collect data on various cross-sections concurrently. Transits provide the same level of accuracy as a laser level, but require at least two surveyors on site. Hand levels have lower accuracy and require at least two surveyors, but are very transportable. For surveys in remote, rugged terrain, hand levels are often the preferred surveying equipment. Stretching a level line across a stream channel and directly measuring vertical distance with a rod is commonly used for cross-section measurement. Errors occur when the line is not level from left to right bank or when the line sags in the middle. Small gauge cable marked at regular intervals with beads eliminates much of the sagging and also eliminates fluttering of tapes due to wind.

The overall method for cross-sectional surveys is initially to define the length of the study reach and to then separate it into sub-reaches/cross-sections. A reach should represent a homogeneous length of stream including similar slope, bed and bank material, bed forms, floodplain/terraces, vegetation and dominant channel processes, so that measurement variance is minimized. A reach length of approximately 30 channel widths provides a relatively homogeneous sample unit in many alluvial channels. Selection of the number of sub-reaches or cross-sections of a given type (pool or riffle; convergent or divergent flow; meander bend or straight) is a function of the percentage of these sub-reach types over the length of the entire reach; however, cross-sections that are spaced closer than three channel widths can result in autocorrelation unless there is a significant, abrupt change in morphology. Systematic sampling of sub-reaches based strictly on multiples of channel width rather than on morphological types may result in over- or under-sampling of specific channel types because of cyclic spacing, especially in alluvial channels (e.g. pool-riffle spacing of approximately seven channel widths; Leopold *et al.* 1964; Hey and Thorne 1986). Once reaches and sub-reaches have been defined, data collection proceeds according to standard surveying techniques (see Harrelson *et al.* 1994). Reach-average values automatically represent a weighted average of the conditions at the surveyed cross-sections.

Measurement of cross-sectional change

Cross-sectional changes are investigated through the acquisition and superimposition of existing topographic surveys (Fig. 11.3). Historical series of cross-sections provide additional information compared with longitudinal profiles, as they allow the investigation of changes in cross-sectional parameters, including width, depth, width-to-depth ratio, and of net erosion or deposition (Murphey and Grissinger 1985; Simon 1992; Rinaldi *et al.* 1997; Agnelli *et al.* 1998).

Problems using historical cross-sections are mainly related to the identification of reliable common points in survey comparisons. Reference water stage elevation can also be a problem, as bankfull stage in historical sections is usually not available. In such cases, changes in the overall cross-section are generally measured, with reference to the maximum water stage contained in the channel. Current and future changes can be investigated by periodic measurements through one of the topographic methods reported previously.

Studies of changes in cross-section should also include surveys and quantification of changes in floodplain elevation and topography. Floodplain surfaces commonly originate as depositional bars within active stream channels. These incipient floodplains build vertically over time as additional sediment is accreted. Vegetation and downed wood provide increased hydraulic roughness, reducing velocities and therefore encouraging more rapid deposition. Areas of deposition can be identified by partial burial of the trunks of woody-riparian vegetation, freshly deposited sediment over organic material (O horizon of the soil profile) or partially buried cultural features. More reliable estimates of deposition rates can be obtained by dating trees using an increment borer and then excavating the tree to the depth of the root collar or root flare (Simon and Hupp 1992). This represents the germination point of the tree. An average deposition rate can then be calculated by dividing the amount of burial by the age of the tree. Event-related deposition can be identified and dated based on even-age-class stands of woody-riparian vegetation. More detailed data can be obtained by placing clay pads in various locations on the floodplain surface and measuring the amount of deposition that occurs on these pads following floodplain inundation.

Planform characterization and measurement

Channel planform refers to the 2D planimetric characterization of the channel morphology. Measurements of channel planform include number of active channels, sinuosity, meander belt width, confinement, meander amplitude and wavelength, radius of curvature and valley width. From these measurements, various parameters related to flood and sediment routing can be calculated and the relative type and abundance of habitat types can be inferred. Furthermore, planform parameters such as sinuosity and braiding indices are often used for channel

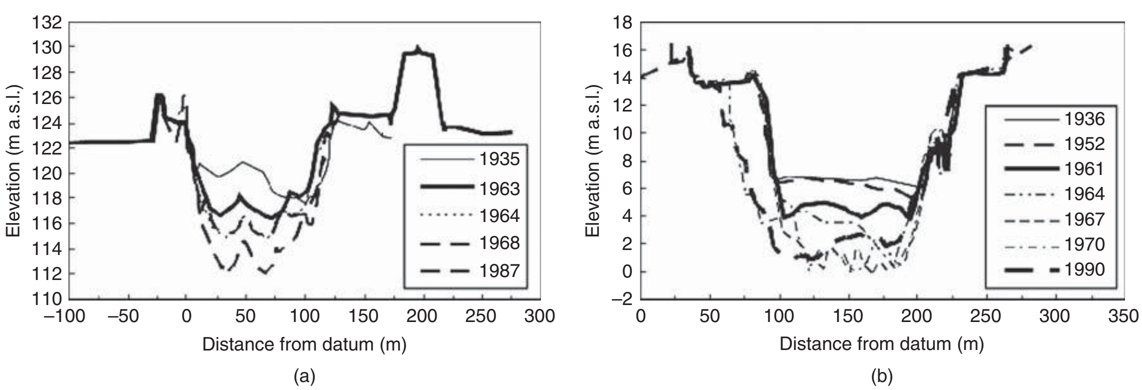


Figure 11.3 Changes in cross-sections: examples from the Arno River. Rinaldi and Simon, 1998. Reproduced with permission of Elsevier.

morphological classification (e.g. Leopold and Wolman 1957; Brice 1975; Rust 1978; Thorne 1997).

Similarly to cross-section parameters, planform parameters have also been used in hydraulic geometry studies and are useful for characterizing channel geometry and for morphological classification. As with other parameters, these values can be combined with longitudinal, cross-sectional and hydraulic data to calculate channel velocity and discharge at various stages, stream power, shear stress and other parameters that are used to quantify channel processes.

Planform characterization

Direct field surveys utilizing distance, bearing and elevation provide a current representation of channel sinuosity and planform and may be preferable to aerial photographs for small streams. Field surveys using total stations are very effective and efficient for stream systems that have minimal vegetative cover or for areas with deciduous foliage. In heavily vegetated areas, field surveys utilizing hand levels, compasses and hip chains or tapes become a more efficient alternative, although the accuracy is greatly reduced. In many instances, GPS can be employed to carry out rapid and precise distributed surveys of planform.

One of the simplest methods for characterizing channel planform and measuring planimetric parameters is the use of maps and aerial photographs. The application of satellite images is increasingly common, but their use is still limited to sufficiently wide rivers, owing to limitations in resolution and therefore in precision of the measurements. Ground-based images acquired by software-controlled digital cameras have recently been used for detailed investigations on changes in morphological planimetric parameters of braided rivers in response to changing discharge and bed evolution (Welber *et al.* 2012).

GIS software is widely applied for georectification of the images and for digitizing channel features useful for planform parameters. Limitations and errors due to georectification and digitizing of channel morphological features have been described previously by several authors (e.g. Gurnell 1997; Winterbottom 2000; Hughes *et al.* 2006).

Accurate channel planform maps allow the measurement of channel width as distance between channel margins. Generally, a series of measurements of channel width along a reach are performed along cross-sections orthogonal to the geometric channel axis, with a spacing of the order of 0.5–2 times the mean width of the reach (Surian *et al.* 2009a). Alternatively, the channel area for a given reach can be measured and then divided by the channel length along the reach. If channel cross-section data have been collected in the field, then remote measurements can be further calibrated. If vegetated islands are present along the reach, two separate measures that include and exclude the islands can be used, depending on the aims of the study.

Sinuosity has been defined in various ways, including (1) channel thalweg length divided by valley length (Leopold and Wolman 1957), (2) channel length divided by valley length (Brice 1984; Schumm 1985), (3) channel length divided by (a) the length of the channel–belt axis (Brice 1964), (b) the meander–belt axis (Alber and Piégay 2011) or (c) the overall planimetric course axis (Malavoi and Bravard 2010) or (4) valley slope divided by channel slope. In all cases, measurements can be carried out with aerial photographs and GIS software on relatively large rivers and with field surveys for smaller streams.

Similarly to sinuosity, many definitions of the braiding index have been proposed (Thorne 1997); the most widely used is the mean number of active (wetted) channels per transect (Ashmore 1991). A review of methods for defining and measuring braiding intensity was presented by Egozi and Ashmore (2008). The braiding index of a reach is obtained as the mean value of a series of transects with longitudinal spacing no less than the channel width and for a reach length no less than 10 times the channel width. A main limitation in measuring the braiding intensity is that it depends on the river stage at the moment of the aerial photograph (Belletti *et al.* 2013). This sensitivity to river stage has been clearly shown by repeated measurements derived from ground-based images over a 2 years study period along two reaches of the Tagliamento River, Italy (Fig. 11.4) (Welber *et al.* 2012). Based on these findings, excessively low or relatively high (around bankfull) river stage conditions should be avoided, as the braiding index in these cases can be largely underestimated (Bertoldi *et al.* 2009; Surian *et al.* 2009a).

Measurement of planform change

Measuring planform changes over time can provide an accurate estimate of system variability and valuable information for potential rates and magnitude of future channel changes. Evidence of meander translation and compression may indicate variable boundary resistance, whereas braiding indicates high sediment loads or sediment pulses. Consistent features that have not migrated up- or downstream over time may represent significant valley controls or gradient breaks.

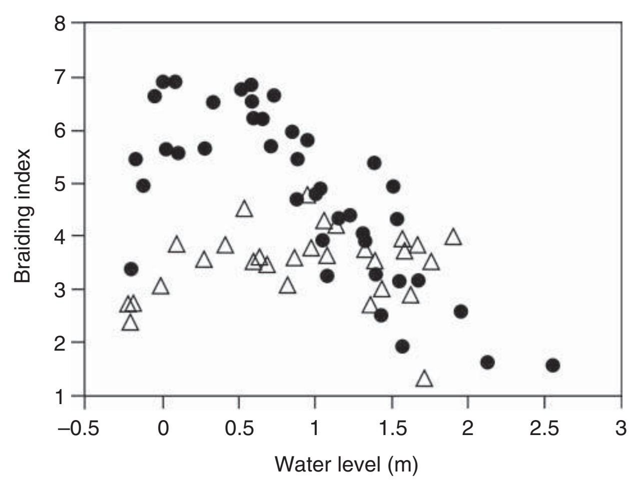


Figure 11.4 Variability of the braiding index with water level along two reaches of the Tagliamento River (Northern Italy) investigated by ground-based images. Symbols (circles and triangles) refer to two different reaches (Comino and Flogogna, respectively), while water level is measured at the Venzone gauging station. From Welber *et al.* (2012).

Maps and aerial photographs are largely used for quantifying changes in channel planform with time. Historical maps provide important information regarding channel position, complexity and simplification prior to the advent of aerial photography (see Chapter 4). Old maps (from the 16th to the 19th century) from archives and from historical studies can only provide qualitative information but are often useful for assessing the channel morphology prior to the main human pressures and to understand better the types and locations of human interventions (e.g. Petts *et al.* 1989). Aerial photographs are generally superior to maps because there is no filter of interpretation (see Chapter 6). Aerial photographs with decadal intervals are ideal for determining planform changes over time. Streams represented on photographs with various scales can be digitized and overlain in a GIS-based system, which allows for direct comparison and analysis. Application of satellite images to the study of morphological changes has progressively increased with the improvement of image resolution, particularly for large rivers (e.g. Thorne *et al.* 1993). Use of maps, aerial photographs and other remotely sensed data allows the analysis of spatio-temporal trends of the planimetric parameters and indices previously defined and identification of temporal changes in channel configuration. For this type of analysis, errors in photograph set measurements deriving from distortions, photographic quality, resolution, colour, scale and other attributes, can significantly affect the interpretation of changes and should be taken into account (e.g. Mount *et al.* 2003; Swanson *et al.* 2011).

Similarly to bed-elevation data, two types of representation can generally be used for planimetric parameters: (i) spatio-temporal distributions by plotting the parameter versus distance downstream for different years; and (ii) temporal trend, by plotting the mean value of the parameter along a given reach versus time. The first type of representation allows the visualization of the spatial variation of a given planimetric parameter and, at the same time, comparison of values at the same position for different years (Fig. 11.5). The second type of representation provides information on the temporal trend, or trajectory, of the parameter (Fig. 11.6a).

The parameter that is frequently used for this analysis is channel width. When comparison among different reaches or rivers is carried out, a dimensionless width, W/W_{max}, is more appropriate, where W is the width measured on the different dates and W_{max} is the maximum width for each reach for the investigated period (Fig. 11.6b) (Rinaldi *et al.* 2008; Surian *et al.* 2009b).

Temporal trends of the sinuosity index can be performed to verify whether changes in morphological pattern occurred (e.g. from sinuous to meandering). Historical trends in braiding intensity are also extremely useful for investigating changes in channel pattern (Gurnell *et al.* 2009). However, for this parameter more uncertainty is involved, as the braiding index depends on the water stage, so measurements should refer to aerial photographs carried out during similar water levels.

Systematic analysis of meander changes, with calculation of erosional and depositional areas or morphometric variations in channel bars, is also used to investigate channel adjustments, instability, mechanisms and propagation of change (e.g. Hooke 2007; Church and Rice 2009).

Three-dimensional characterization and measurement

Recent technological advances provide new possibilities for characterizing river morphology through 3D acquisition and representation of topographic data at high spatial resolutions (Lane and Chandler 2003; Milan *et al.* 2007; Marcus and Fonstad 2008; Wheaton *et al.* 2010). Digital elevation models (DEMs) built from such surveys have many objectives and applications, including hydrodynamic and morphodynamic modelling, characterization of river features and habitats and quantification of morphological changes and sediment budgets (see the next section).

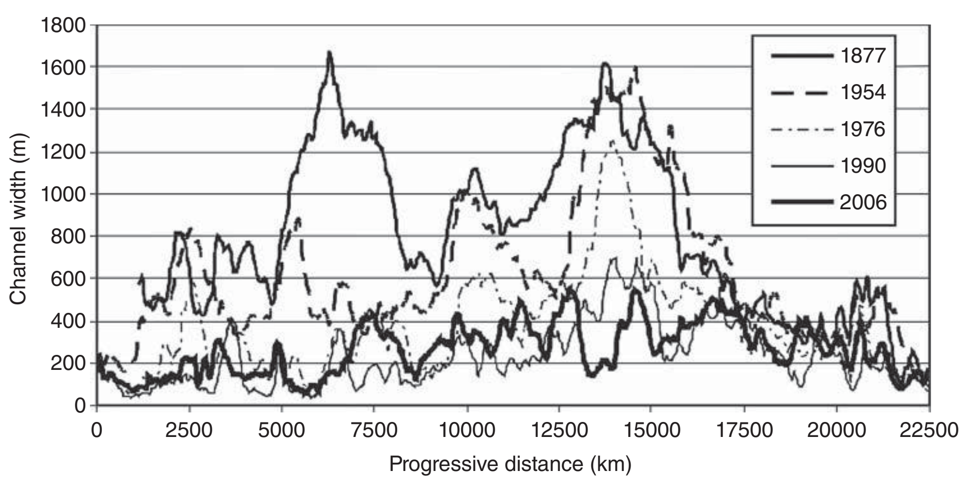


Figure 11.5 Spatio-temporal changes in channel width: example from the Trebbia River (Northern Italy). Source: Pellegrini *et al*, 2008. Reproduced with permission of Il Quaternario.

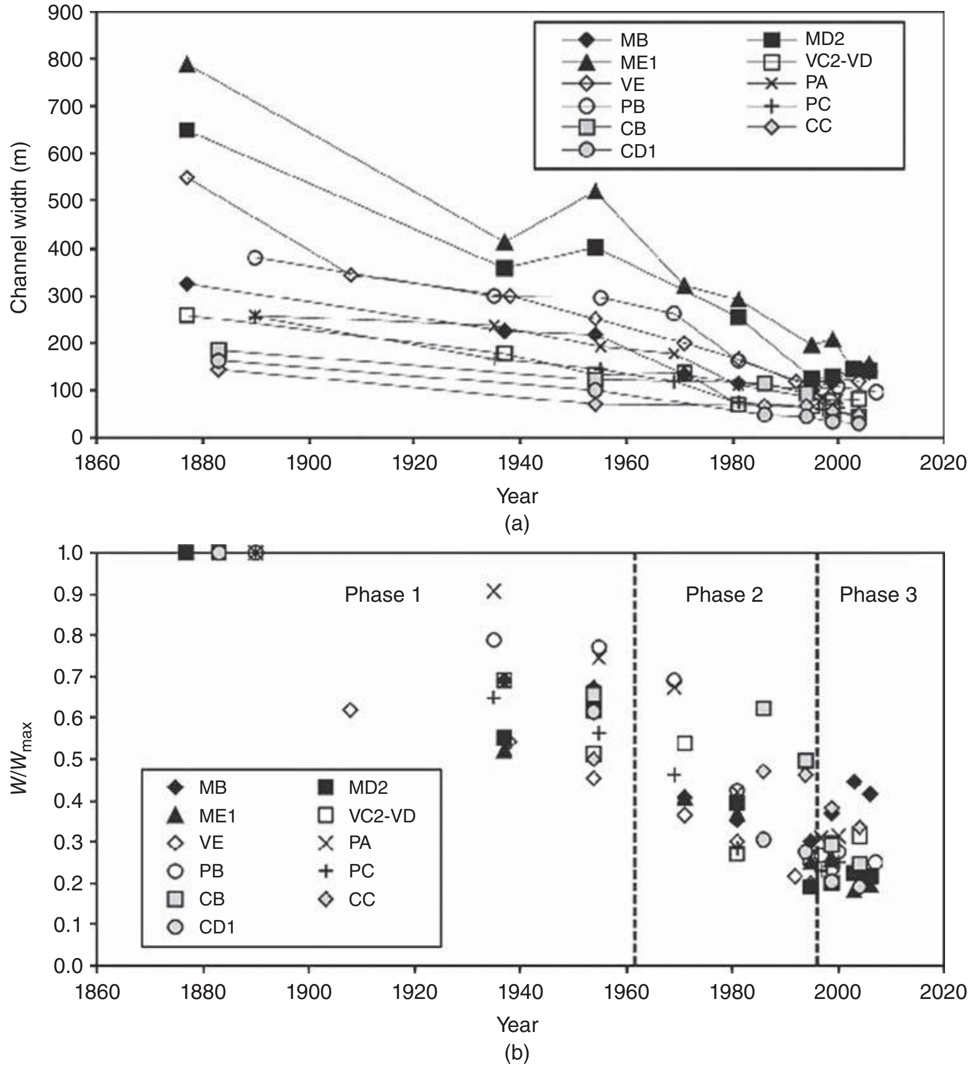


Figure 11.6 Changes in channel width: examples from some Apenninic rivers (central–northern Italy). (a) Temporal trends of channel width along selected alluvial reaches of four study rivers (MB, …, CC); (b) changes of dimensionless channel width, expressed as W/W_{max}, where W is the width measured on the different dates and W_{max} is the maximum width of the investigated period. Source: Rinaldi *et al*, 2008. Reproduced with permission of Il Quaternario.

Three-dimensional characterization

DEMs of the river bed, adjacent surfaces or portions of the channel (e.g. riverbanks) can be obtained by different ground-based, boat-based and remotely sensed surveying technologies, including GPS (Brasington *et al.* 2000), aerial photogrammetry (Westaway *et al*. 2001; Hicks *et al.* 2008), terrestrial photogrammetry (Pyle *et al.* 1997), airborne LiDAR (Cavalli *et al*. 2008) and terrestrial laser scanning (Lichti *et al.* 2002; Milan *et al*. 2007). Active remote-sensing instruments, such as terrestrial laser scanners, offer a significant improvement in the speed, accuracy, density, volume and spatial coverage of terrain data that can be collected on river banks (e.g. O'Neal and Pizzuto 2011). Small radio-controlled motorized vehicles flying at low altitude can be used to obtain very high-resolution images and to study both channel water depth and gravel bar geometry (Lejot *et al*. 2007).

Measurement of three-dimensional channel change

Advances in surveying technologies (as described above) make possible the study of channel changes, quantification of erosion and deposition and estimation of sediment budgets through repeated topographic surveys. DEMs obtained by various survey techniques (GPS, total station, photogrammetry, laser scanning) can be used to produce DEM of difference (DoD) maps and estimate the net change in sediment storage. Sediment budget and estimation of sediment transport can be obtained by quantifying volumetric changes and applying the morphological method (e.g. McLean and Church 1999; Brewer and Passmore 2002).

In general, the various surveying techniques employed for such applications have accuracies in surface elevation that can vary from ±0.02 to ±1 m, which can be of the same order of magnitude as channel changes. Consequently, morphological studies and sediment budgets must account for uncertainty in DoD applications (Wheaton *et al.* 2010). Laser scanners, both airborne LiDAR and terrestrial, appear to be the most promising techniques for building the highest resolution DEMs. For example, repeated field-based laser-scanner surveys can be used at both the site and reach scale to produce time series of centimetre to decimetre-scale DEMs of channel surfaces subject to rapid changes in form (Milan *et al.* 2007; O'Neal and Pizzuto 2011).

Bed and bank characterization and measurement

Bed and bank material sampling and characterization are normally carried out during a study to characterize broader channel forms and processes. Measurements of bed and bank material generally include particle size distribution, stratification/sorting, patchiness and percentage fines. From these measurements, indices such as degree of bed armouring, relative particle mobility or erodibility coefficients can be calculated.

Bed sediment size is fundamental for estimating bed roughness and sediment transport capacity and for habitat characterization. The occurrence of bed armouring can provide information on watershed sediment supply and hydrological conditions (e.g. Hassan *et al.* 2006), and also on the existence of alterations in water and sediment regimes (e.g. dams). Sediment sampling at multiple stream sites provides information on bed variability and fining, in relation to the supply of tributaries and variable transport conditions (e.g. Knighton 1982; Rice 1998). Similarly, bank sediment characterization is useful for bank erodibility and analysis of downstream controls on bank stability, roughness, quantification of sediment supply from lateral processes and characterization of riparian habitats. For bank material sampling and testing, a distinction must be made between bank surface sediment and internal bank material because characteristics of the former are related to hydraulic processes, whereas the latter are related to geotechnical processes.

Bed and bank characterization

Bed material samples (both armour and sub-armour if present) should be obtained from each of the cross-sections selected in the reach and should represent the morphological features present. Methods to sample boundary sediments vary by the size class of sediments and their position along the channel boundary. Particle-size characteristics of bed-material samples are used to calculate the shear stress required to erode non-cohesive beds (see Section 11.3). Sampling of coarse-grained bed material is covered in detail in Chapter 13.

For cohesive (silt/clay) streambeds, different sampling techniques are used to estimate critical shear stresses and erodibility coefficients, including a submerged jet-test device (Hanson 1990, 1991; Hanson and Simon 2001; Simon *et al.* 2011). In the absence of jet-test measurements, the percentage clay or plasticity index of cohesive streambeds can be used as a measure of the relative resistance to erosion, with higher values indicating greater resistance.

In studies specific to riverbank stability and failures, bank characteristics need to be measured in detail, which can be achieved by a careful field survey aimed at obtaining an inventory of current bank geometry, materials, stratigraphy, profiles, erosion processes and failure mechanisms. Field reconnaissance forms (Thorne 1998) should include specific sections focusing on bank conditions and processes.

Basic geometric parameters include bank height and mean slope that can be easily measured by pocket rod and clinometer. Classification of bank type (non-cohesive, cohesive, composite or layered) and characterization of bank sediments is important, together with interpretation of dominant processes and identification of relic tension cracks, failure surfaces, bank toe and geometry of cantilever blocks.

Samples of bank material are collected to determine the erosion resistance due to (i) hydraulic forces, which erode bank surfaces by the shear of flowing water, and (ii) gravitational (geotechnical) forces, which erode banks through mass-failure mechanisms. Bank material samples should be collected by bulk sampling if possible and analysed to determine the particle-size distribution for potential use in calculating critical shear stress and the percentage of major size classes (i.e. sand, silt and clay) for use in selecting 'typical' values of geotechnical properties (Simon *et al.* 2011). Because streambanks also erode by mass failure, they need to be sampled/tested for their geotechnical properties. These parameters include effective cohesion, angle of internal friction, pore-water pressure and bulk unit weight. Total cohesion and friction-angle data can be obtained from standard laboratory testing (triaxial shear or unconfined compression tests) or by in situ testing with a borehole shear test (BST) device (Lohnes and Handy 1968; Lutenegger and Hallberg 1981; Thorne *et al.* 1981; Little *et al.* 1982; Simon 1989a).

Since shear strength (total cohesion) varies with moisture content, measurements of pore-water pressure are made in conjunction with the BST at all test depths to obtain values of effective cohesion. These can be obtained by extracting a core from the appropriate depth with a hammer sampler. A portable piezometer/tensiometer is then inserted into the core to determine the magnitude of pore-water pressure or matric suction (negative pore-water pressure) (Rinaldi and Casagli 1999; Simon *et al.* 1999). Effective cohesion (c') is then calculated as the difference between total cohesion and cohesion due to matric suction.

Measurement of bed and bank change

Measurement of changes in bed elevation can be achieved by repetition of topographic surveys using the same techniques described previously for longitudinal profile and cross-sectional changes.

As an additional method to measure more specifically short-term changes in the elevation of non-cohesive channel beds, scour chains can be installed (Laronne *et al.* 1994). Chains are anchored to a pin placed horizontally below the estimated maximum depth of scour, extend vertically upwards and then drape over the bed surface. The elevation of the bottom of the chain and the bed surface are surveyed and the length of chain exposed is measured. Scour chains are inspected and measured after peak-flow events. The exact location of the scour chain should be carefully noted, because finding the scour chain after a large flow event is very labour intensive and can be extremely difficult. During a sediment-transporting event, the material around the chain may be scoured, causing the chain to lie over against the remaining bed material. This level indicates the depth of scour for the event (Lisle and Eads 1991). If a scour chain is left in place through numerous events, it is not possible to determine which event caused the greatest degree of scour, so it is essential to locate, measure and replace the scour chain after every significant event if the purpose is to determine average scour conditions for various sized events.

A comprehensive review of the methods used to observe bank erosion was provided by Lawler (1993), who highlighted how different methods can be appropriate depending on the time-scale of investigation. Techniques used for investigation of cross-section or planform changes are often also applicable to bank erosion. Recent technological developments, such as digital photogrammetry and laser scanning (e.g. Pyle *et al.* 1997; O'Neal and Pizzuto 2011), can provide the opportunity to define river bank topography at unprecedented spatial resolution (surveys with point densities of ~100 points across a bank face are readily obtainable using terrestrial laser scanning) and accuracy (~70 mm). Bank erosion can then be quantified using the survey data to construct DEMs for time intervals and differencing to establish net change.

A method more specifically employed for fluvial erosion at the bank-toe consists in using a network of erosion pins spaced longitudinally along the bank (Simon *et al.* 1999; Stott 2005). Each set of pins is made up of pieces of rebar inserted horizontally into the bank-toe region and displaced vertically. The length of exposure for each pin is measured after runoff events or at a frequency conducive to the temporal scope of the study. Based on measurements of protrusion lengths made by different operators on the same day, erosion-pin data are accurate to within ±5 mm (Simon *et al.* 1999). Estimates of the change in length between visits are accurate, therefore, to within 1 cm. However, additional systematic errors and uncertainty may be introduced through the effects of (i) turbulent scour around the tip of the pins, (ii) disturbance of the bank-material fabric during insertion of the pins and (iii) erosion in excess of the pin length and loss of the pin.

Application of topographic methods and erosion pins can provide measurements of bank erosion following, at best, a single flow event, but are unable to provide a continuous monitoring of erosion processes. To address this limitation, new quasi-continuous bank-erosion sensors based on the use of photo-electronic cells (PEEPs) (Lawler 1993; Lawler *et al.* 1997) and thermal consonance timing (TCT) (e.g. Lawler 2005, 2007) have been developed, although they have not yet been widely deployed, as they are susceptible to breakage from impact of sediment, debris or mass-wasting events.

Channel-forming discharge characterization and measurement

To estimate fluvial responses for various flow events, it is necessary to determine which events are significant in the context of sediment transport for a specific study, project or design. For flood-control works, discharges such as the Q_{10}, Q_{25}, Q_{50} and Q_{100} are typically evaluated. In the realm of fluvial geomorphology, more frequent discharges are often of greater interest, although it is still important to evaluate the potential geomorphic response for a large event. In recent years, flows such as the 'channel-forming,' 'bankfull,' 'dominant' or 'effective' discharge have been widely used and referenced. It should be stressed that the 'effective discharge' is a concept that probably represents a range of flows and is not to be confused with 'bankfull discharge' and 'dominant discharge.' So as to avoid confusion, definitions are given as follows. 'Channel-forming' or 'dominant' discharge is intended as a *theoretical* single recurring or steady discharge that, given enough time, would produce the same morphology and dimensions as produced by the actual flow regime (Inglis 1949). There are at least three approaches to determining the channel-forming discharge (Biedenharn *et al.* 2001; Shields *et al.* 2003; Soar and Thorne 2012): (1) bankfull discharge; (2) discharge of a given recurrence interval; and (3) effective discharge.

1 *Bankfull discharge:* For unconfined, non-incised, alluvial streams, the maximum discharge that can be contained within the channel without overtopping the banks and flowing onto the active floodplain. In the streams in which the seminal studies were conducted (primarily US Geological Survey gauging stations in humid-climate, snowmelt-dominated, typically large perennial streams), bankfull discharge corresponds to the flow that occurs, on average, every 1–2 years (Leopold *et al.* 1964). Leopold *et al.* (1964) specifically state: 'There is a remarkable similarity in the frequency of bankfull stage on a variety of rivers in diverse physiographic settings and differing greatly in size. The recurrence interval of the bankfull stage appears to be in the range of 1 to 2 years, although some localities studied diverge greatly from this value. At stations where the flood plain is clearly defined and its elevation accurately known, the recurrence interval is closer to 1 than to 2 years. In general, a value of 1.5 years seems a good average'. Castro and Jackson (2001) found that for streams in the Pacific Northwest of the United States, variations in recurrence intervals could be explained by ecoregion, with the most important factor being climate. More humid regions, such as the Coast Range of Oregon and Washington, displayed recurrence intervals of 1 year or more frequent, while the Intermountain West was much closer to the 1.5 years average described by Leopold *et al.* (1964) (Castro and Jackson 2001).The concept of the

bankfull channel as shaped by the 1.5 years flow is easily comprehended and has been widely adopted by scientists in allied fields and restoration practitioners, but in climates with more variable hydrology, less frequent floods have a greater role in shaping the channel (Wolman and Gerson 1978). Williams (1978a) evaluated the field identification method to determine bankfull (along with 10 other methods) and concluded that there was a significant range in values (from 1 to 32 years). However, Williams (1978a) did indicate that the valley flat stations used in his analysis (28 of the 64 stations) 'generally are of uncertain significance because of the possibility that some channels may be incised', which would result in larger recurrence interval values. Williams (1978a) also identified a relationship between stream slope and increasing recurrence intervals: 'very generally, the recurrence interval T is greater (longer periods between bank-full flows) as slope steepens'.

2 *Discharge of a given recurrence interval:* The flow with a given return interval is often assumed as the channel-forming discharge, normally $Q_{1.5}$ or Q_2 (i.e. 1.5 or 2 years return interval discharge, respectively), but the caveats noted above apply.

3 *Effective discharge:* The discharge, or range of discharges, that transports the largest proportion of the annual sediment load over the long term (Wolman and Miller 1960; Andrews 1980). Although originally defined for suspended-sediment load, subsequent applications have used bed load, bed-material load and total load.

Bankfull discharge

Indicators of bankfull stage can be estimated analytically or based on field observations. In stable, 'natural' streams, the best indicator of bankfull stage is often the active floodplain surface. This surface, however, is not always identifiable or even present, particularly in steep, cobble-boulder streams and along braided, incised or aggraded channels. In the absence of a well-defined floodplain surface, other indicators, the importance of which will depend on the specific fluvial environment, are useful. Along braided streams or streams containing bars, the top of the bar surface (the proto-floodplain), particularly if it supports woody vegetation, is often a good indicator of bankfull stage, although it represents a minimum level for bankfull. In incised channels, where the previous floodplain surface has become a terrace, the bankfull stage can be identified as the lower-most limit of establishing woody-riparian vegetation. Williams (1978a) and Harrelson *et al*. (1994) list additional useful indicators, although they should be used with caution and accepted only if there are other lines of evidence.

Field-based observations should, if possible, be verified with stream-gauge data (Leopold 1994). Stage-discharge relations established at gauging stations are available in the United States from the US Geological Survey and in other countries from those agencies responsible for monitoring water resources. If the stage-discharge relation abruptly changes to a flatter slope at higher discharges, this represents the stage at which flow spreads out across the floodplain or braid-plain surface. If the elevation of this stage is in agreement with surveyed, field-based observations of bankfull indicators, one can be reasonably confident in the selection of the bankfull stage. For US Geological Survey gauging stations, gauge location is often based on relative channel stability, hydraulic control, single-thread channels and other attributes that result in more consistent and reliable measurements of flow and as such may represent a biased sample with regard to channel morphology (Castro 1996).

Discharge of a given recurrence interval

If gauge data are available, the flow with a given return interval is often assumed to be the channel-forming discharge. In large, perennial streams, the channel-forming discharge in stable channels often corresponds to a flood recurrence interval of approximately 1–2 years (Leopold *et al.* 1964; Andrews 1980; Castro and Jackson 2001; Simon *et al*. 2004), with 1.5 or 2 years the most frequently used values. However, as noted above, the actual range of return intervals for bankfull discharge is much wider (Williams 1978a), and in streams with more variable hydrology or with steeper slopes, channel form may be dominated more by infrequent events (Wolman and Gerson 1978).

Effective discharge

A convenient means of estimating the long-term potential for sediment transport is to compare critical and available shear stresses for the effective discharge. Again, for large, perennial streams, the effective discharge commonly occurs, on average, about every 1–2 years (Simon *et al.* 2004), but as noted by Wolman and Miller (1960, p. 60), 'The … more variable the regimen of flow of the stream, the larger the percentage of total sediment load which is likely to be carried by infrequent flows'. In Mediterranean-climate California, of the roughly 57.6 million tonnes of sediment load measured on the Santa Clara River at Montalvo from 1968 to 1975, 55% was moved in only 2 days during the 1969 flood (Williams 1979), so in a river such as this, the effective discharge is not the $Q_{1.5}$, but rather the largest discharge on record. To determine the actual effective discharge, an established sediment-transport relation (concentration versus flow) and a flow-frequency distribution are required. A comprehensive guide to effective discharge calculation is reported by Biedenharn *et al.* (2001). This involves a three-step process:

1 construct a frequency distribution (histogram for discharge);

2 construct a sediment-transport rating relation; and

3 integrate the two relations by multiplying the sediment-transport rate for a specific discharge class by the frequency of occurrence for that discharge, with the maximum product being the effective discharge.

Mean-daily flows are often used with instantaneous values of suspended sediment concentration because these data are readily available. The minimum period of record should be at least 10 years. Except for large rivers, this approach is biased towards low-flow conditions because short-duration peak discharges are neglected. A better approach is to use shorter termed flow data, such as those corresponding to the 15 minute stage data. If sediment data are not available, bed-material load transport rates can be derived from a variety of transport functions. Procedures and examples are provided in Stevens and Yang (1989) and Andrews and Nankervis (1995). Regardless of the type of discharge data used, data are ranked and then subdivided into 25–33 classes (Yevjevich 1972). Subdividing classes using an arithmetic distribution often results in the majority of flows falling into the lowest discharge class. To overcome this problem, a logarithmic distribution is used. The effective discharge can then be calculated as the discharge class that has the maximum sediment concentration/discharge product for the classed flow-frequency data.

11.3 Interpretation and analysis

Characterization and measurement of channel form provide a context for analysis and interpretation of present and future channel morphologies and are often a central theme in studies that strive to identify and determine the magnitude and extent of channel change. Because streams are open systems, an alluvial channel has the ability to adjust to altered environmental conditions.

Adjustment processes that can affect entire fluvial systems include channel degradation and aggradation, lateral channel migration, channel widening or narrowing, channel avulsion and changes in the quantity and character of the sediment load. These processes differ from short-term, event-related localized processes such as scour and fill, which can be limited in magnitude and also in temporal and spatial scale.

Scour and fill in a streambed over the course of a storm hydrograph, although representing streambed mobility, do not necessarily indicate instability because the short time period of the event is not indicative of rapid, progressive change. Slow, progressive erosion of a meander bend with concomitant deposition on the opposite point bar, which maintains an average channel width over time, also does not indicate instability. Meander migration of an alluvial channel is expected over long periods; again, potential instability is not inherent in the change, but rather a result of the rate of change.

The previous examples highlight the importance of time-scales in interpretation and analysis of channel form. Even the dependence of variables can change as a function of the time-scale applied. Schumm and Lichty (1965) showed how variables describing channel form are indeterminate over geological time, dependent over medium time-scales and independent over short time-scales.

Concerning spatial scale, analysis of changes in channel form requires that the investigator determine whether processes are localized disturbances or system-wide adjustments. It is difficult to differentiate between localized and system-wide processes without extending the investigation upstream and downstream of a particular stream reach. Similar channel forms can be the result of dissimilar causes and, because channel adjustments migrate over time and space and may affect previously undisturbed reaches, it is essential to identify properly the cause of channel change rather than the symptoms. To determine channel stability or simply to quantify channel processes, measurements of channel changes are necessary.

The purpose of this section is to provide a guide for analytical techniques related to alluvial channel form that are central to understanding aspects of alluvial channel behaviour. This is generally accomplished by considering those factors that directly control the balance or imbalance between applied forces and boundary resistance. Generally, if force and resistance are balanced, the stream neither rapidly erodes nor fills and is capable of transporting the sediment load delivered from upstream reaches. This balance indicates a stability of channel dimensions and can be expressed mathematically as the stream power proportionality (Gilbert 1914; Lane 1955):

$$QS_b \propto Q_s D_{50} \tag{11.1}$$

where Q = discharge, S_b = bed slope, Q_s = bed-material discharge and D_{50} = median grain size of bed material, indicating that 50% of the bed material is finer.

Equation 11.1 indicates that if available stream power were augmented by an increase in the discharge or the gradient of the stream, there would be an excess amount of stream power relative to the discharge of bed-material sediment delivered from upstream. Additional sediment would be eroded from the channel boundary resulting in (i) an increase in bed-material discharge to an amount commensurate with the heightened stream power and (ii) a decrease in channel gradient and, consequently, stream power as the elevation of the channel bed is lowered. A similar response would be expected from a decrease in the erosional resistance of the channel boundary or a decrease in the size of bed-material sediment (assuming the bed is not cohesive). In contrast, a decrease in available stream power or an increase in the size or discharge of bed-material sediment would lead to aggradation of the channel bed. Aggrading or degrading channels represent end members on a continuum where vertical stability is represented at the centre point.

The conceptual and semiquantitative relation provided by eqn. 11.1 provides only limited insight into the type and hierarchy of adjustment processes. Excess stream power can erode additional sediment from the channel boundary; however, eqn. 11.1 does not indicate where the erosion will occur and, therefore, how channel form might change. Identifying instream sediment sources in this case becomes a matter of determining the relative resistance of the bed and bank material to the applied forces imposed by the flow and gravity. For a sand-bedded stream with cohesive banks, an initial adjustment might involve streambed incision because of low critical shear stresses, higher applied shear stresses on the bed than on the bank-toe and more frequent exposure to hydraulic shear than adjacent streambanks. Conversely, if we assume that the streambed is highly resistant, composed of cohesive clays, bedrock or large particles such as cobbles or boulders and that the bank-toe is composed of significantly weaker materials, we could expect bank erosion to be the initial adjustment.

A review of some of the techniques that have proven useful in analysing these processes is presented in this section. Once the dominant processes have been identified and the appropriate physical data collected, various empirical and numerical methods are available to quantify subsequent changes in channel form and to estimate future channel configurations (Thorne *et al.* 1981; Schumm *et al.* 1984; Hey and Thorne 1986; Simon and Hupp 1986, 1992; Chang 1988; Thorne and Osman 1988; Molinas 1989; Lohnes 1991; Simon and Downs 1995; Langendoen 2000). These studies include 'regime' and other empirical methods in addition to numerical simulation models. There is insufficient space here to review all of these in detail and readers are directed, therefore, to these publications and to the following chapters of this volume: Chapter 17, Models in fluvial geomorphology; Chapter 18, Modelling flow, sediment transport and morphodynamics in rivers; and Chapter 19: Modelling fluvial morphodynamics.

Empirical methods

Empirical methods refer to techniques that rely on relations developed from measurements and observations in the field or laboratory, which may or may not be physically based.

Quantifying stable channel dimensions

Numerous empirical methods are available with which to estimate 'stable' channel dimensions. Leopold and Maddock (1953) derived relations between mean velocity, mean depth and water-surface width as a function of discharge (see Chapter 17 for details). Collectively termed 'hydraulic geometry', these relations can be expressed as 'at-a-station' (for a single cross-section) or as 'downstream' relations. For the 'downstream' relations, the bankfull or mean annual discharge is used (Leopold *et al.* 1964). Exponents of the hydraulic geometry equations vary by region due to differences in climate, rainfall–runoff relations and, specifically, by the type and resistance of boundary sediments (Leopold and Maddock 1953; Leopold *et al.* 1964; Williams 1978b; Castro and Jackson 2001). Results from these equations should be used with extreme caution for channel design because of the uncertainty in regression estimates even with high r^2 values.

Similar empirical procedures termed 'regime methods' developed by engineers in studies of irrigation channels generally rely on three formulas to describe a stable width, depth and slope (Lacey 1930, 1958). Generally, bankfull or channel-forming discharge is used to represent the flow regime. Blench (1952, 1970) modified Lacey's approach by accounting for differences due to variability in bank materials and Simons and Albertson (1960) and Hey and Thorne (1986) allowed for channels other than those composed of sand beds and cohesive banks.

Quantifying longitudinal profile and bed elevation changes

In unstable channels, bed elevation with time (years) can be described by non-linear functions, where change or response to a disturbance occurs rapidly at first and then slows and becomes asymptotic. Plotting of bed elevations with time permits the evaluation of bed-level adjustment trends and indicates whether the major phase of degradation or aggradation has passed or is ongoing. Various mathematical forms of this function, including exponential, power and hyperbolic, have been used to characterize bed-level adjustment at a site with time and to predict future bed elevations (Graf 1977; Williams and Wolman 1984; Simon and Hupp 1986; Simon 1989a, 1992; Wilson and Turnipseed 1993, 1994). Extensive studies of bed-level adjustment in streams representing a wide range of bed material sizes have shown that the power and exponential functions accurately describe upstream degradation and downstream aggradation with time (Simon 1989a, 1992). An exponential function converges to an asymptote and is preferable (H. Jobson, US Geological Survey, personal communication, 1992); however, the power function is easier to use:

$$E = at^b \tag{11.2}$$

where a is a coefficient, determined by regression, representing the premodified elevation of the channel bed, t = time since the beginning of the adjustment process (years), where $t_0 = 1.0$ (year prior to onset of the adjustment process), and b is a dimensionless exponent, determined by regression and indicative of the non-linear rate of channel-bed change (negative for degradation and positive for aggradation).

The dimensionless form of the exponential equation is (Simon 1992; Simon and Rinaldi 2000):

$$\frac{z}{z_0} = a + b\mathrm{e}^{-kt} \tag{11.3}$$

where z = elevation of the channel bed (at time t), z_0 = elevation of the channel bed at t_0, a is a dimensionless coefficient determined by regression and equal to the dimensionless elevation (z/z_0) when eqn. 11.3 becomes asymptotic, $a > 1$ = aggradation, $a < 1$ = degradation, b is a dimensionless coefficient determined by regression and equal to the total change in the dimensionless elevation (z/z_0) when eqn. 11.3 becomes asymptotic, k is a coefficient determined by regression, indicative of the rate of change on the channel bed per unit time, and t is the time since the year prior to the onset of the adjustment process (years, $t_0 = 0$).

Future elevations of the channel bed can be estimated by fitting eqn. 11.2 or 11.3 to bed elevations and by solving for the time period of interest. Either equation provides acceptable results, depending on the statistical significance of the fitted relation. The predisturbed bed elevation, obtained from field survey, is required along with at least one other bed elevation from a different time period. Degradation and aggradation curves for

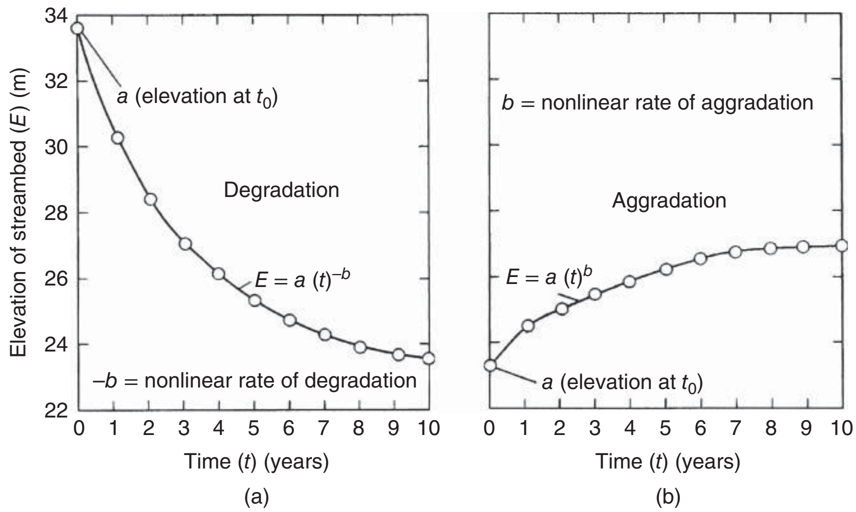


Figure 11.7 Idealized graphs of fitting power functions to (a) degradation and (b) aggradation trends.

the same site are fitted separately (Fig. 11.7). For degrading sites, this method will provide projected minimum channel elevations when the value of t becomes large and, by subtracting this result from the floodplain elevation, will provide projected maximum bank heights. A range of bed adjustment trends can be estimated by using different starting dates when the initial timing of bed-level change is unknown (Fig. 11.8).

The longitudinal distribution of b values (from eqn. 11.2) or a values (from eqn. 11.3) can be used as an empirical model of bed-level adjustment, provided that there are data from enough sites to establish a relation with distance along the channel or river system. An example using eqn. 11.2 is provided for the Obion River system, West Tennessee (Fig. 11.9) and using eqn. 11.3 for West Tarkio Creek, Iowa and Missouri (Fig. 11.10). With knowledge of t_0, b values can be interpolated for unsurveyed sites that can be used to obtain estimates of bed-level change with time. For channels downstream from dams, the shape of the curve in Fig. 11.9 would be similar but reversed; maximum amounts of degradation (minimum b values) occur immediately downstream of dams and attenuate non-linearly with distance further downstream (Williams and Wolman 1984). Once the minimum bed elevation has been obtained using eqn. 11.2 or 11.3, that elevation can be substituted back into the equation and used as the starting elevation at a new t_0 for the 'secondary' aggradation phase (Fig. 11.7b).

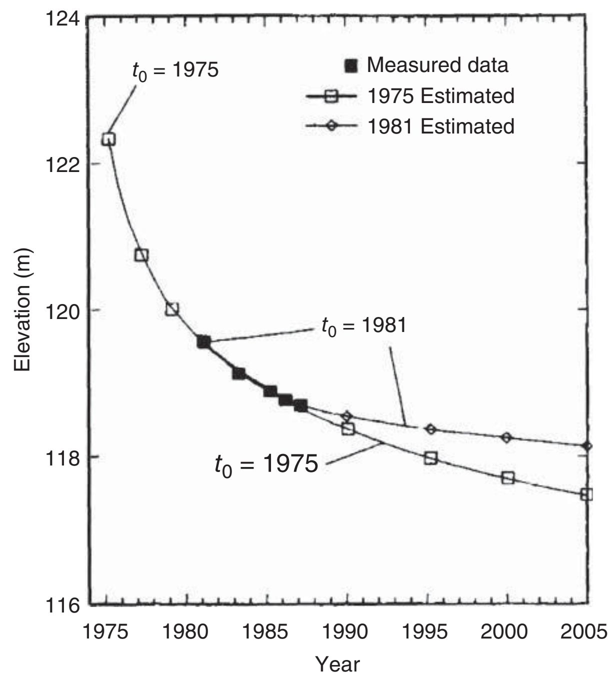


Figure 11.8 Method to predict future bed-level elevations using power functions (eqn. 11.2) and if the timing of the start of the adjustment process (t_0) is unknown.

Quantifying channel width changes and bank erosion

Estimates of potential channel widening on streams with unstable banks can be obtained by noting the angle of the low-bank surface, which indicates renewed stability because of the establishment of supporting woody vegetation. This stable low-bank surface, termed the 'slough line' (Simon and Hupp 1986), is formed as bank angles recede through successive failures and is capped with fluvially reworked material. The horizontal distance between the intersection of the projected slough-line angle with the floodplain surface and the present top bank is the minimum estimated widening for one side of the channel (Simon and Hupp 1992). Various geotechnically based methods of estimating a 'temporary' or 'ultimate' angle of stability have been advanced; however, these methods ignore the cohesion component of shear strength because, in many cases, it is assumed to be zero (Skempton 1953; Carson and Kirkby 1972).

Similarly to bed elevation, estimates of potential channel widening can be made using measured channel width data

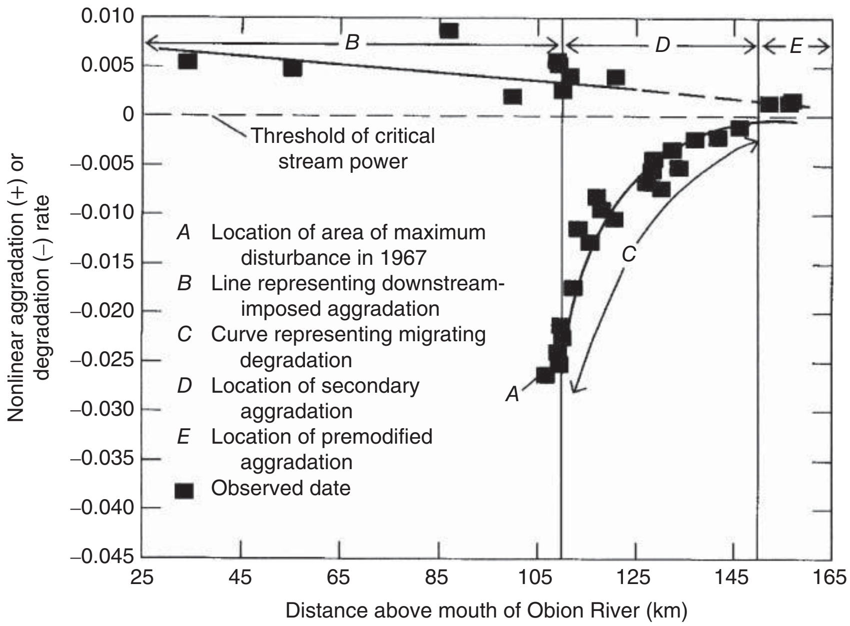


Figure 11.9 Empirical model of bed-level response for the Obion River system, western Tennessee, based on fitting power function (eqn. 11.2) to time series bed-level data. Positive *b* values represent aggradation and negative *b* values represent degradation.

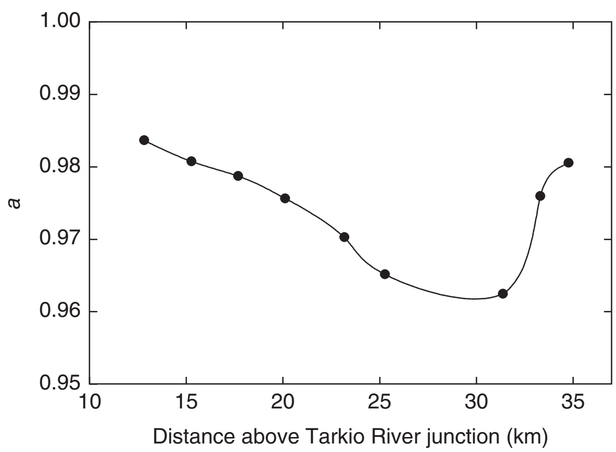


Figure 11.10 Empirical model of bed-level responses for West Tarkio Creek, Iowa and Missouri, based on fitting dimensionless exponential function (eqn. 11.3) to time series bed-level data; *a* values less than 1.0 represent degradation, with the smaller absolute values representing more severe degradation.

over a period of years and then fitting a non-linear function to the data. Williams and Wolman (1984) used a dimensionless hyperbolic function of the form to estimate channel widening downstream from dams. Wilson and Turnipseed (1994) used a power function to describe channel widening after channelization and to estimate potential channel widening in the loess area of northern Mississippi, United States.

Empirical predictions of potential bank erosion zones are generally based on quantification of rates of past bank retreat by aerial photographs and maps. This method is usually employed for reconstructing the erodible river corridor (e.g. Piégay *et al.* 2005).

Another example of an empirical method applied to bank changes is the quantification of parameters related to lateral mobility of single-thread, sinuous to meandering channels. It is known that the depth of pool scour in bendways is related to the geometry of the bend. Empirical relations for bendway scour analysis have been developed based on field data, consisting of prediction of maximum scour depth as a function of the radius of curvature of the bend and channel width (Thorne 1997). Similar empirical relations have been obtained between the ratio of radius of curvature to channel width and the caving rate at the bend, showing that maximum caving rates occur when the radius of curvature is 2–3 times the channel width (Biedenharn *et al.* 1989).

Deterministic methods

Deterministic methods refer to numerical techniques that rely on physically based field and laboratory measurements of the variables that control channel processes. The justification for this approach is fundamental to the science of fluvial geomorphology: that is, the acceleration due to gravity (g) is essentially constant on this planet and, therefore, the physics of erosion, sediment transport and deposition are the same, regardless of the hydro-physiographic province, stream type (Rosgen 1996) or river style (Brierley and Fryirs 2005). These methods require appropriate identification of the active/dominant processes and application of those equations that describe the force and resistance mechanisms for those particular processes.

Stable channel-dimension modelling

Limitations related to empirical methods for estimating stable channel geometry have led many authors to develop approaches that are strongly based on physical processes for determining channel dimensions. Examples are the analytical approaches used in the 'regime' theory, such as extremal hypothesis approaches and tractive force methods. For details on these methods, readers are directed to Chapter 17, Models in fluvial geomorphology.

Longitudinal profile and bed elevation change modelling

Initial modifications to channel form are often manifest by adjustments in profile through erosion or deposition of streambed materials. In the simplest terms, we can conceptualize vertical stability as defined by Mackin (1948), where aggradation or degradation does not occur over a period of years. In physical terms, this indicates that just enough bed material is supplied from upstream relative to the available stream power (eqn. 11.1) or shear stress over a range of flows. Average boundary shear stress (τ_0) is the drag exerted by the flow on the bed and is defined as

$$\tau_0 = \gamma_w R S_b \tag{11.4}$$

where γ_w = unit weight of water (N m^{-3}) and R = hydraulic radius (area/wetted perimeter) (m). The resistance of non-cohesive materials is a function of bed roughness and particle size (weight) and is expressed in terms of a dimensionless shear stress (Shields 1936):

$$\tau *= \frac{\tau_0}{(\rho_s - \rho_w)gD} \tag{11.5}$$

where $\tau *=$ dimensionless shear stress, ρ_s = sediment density (kg m^{-3}), ρ_w = water density (kg m^{-3}), g = gravitational acceleration (m s^{-2}); and D = characteristic particle diameter (m).

Non-cohesive materials. The average boundary shear stress (eqn. 11.4) for a range of flows can be compared to a calculated critical boundary-shear stress (eqn. 11.5) to identify those flows where excess shear stress (erosion) is likely to occur. The Shields criterion is invoked to then calculate the equivalent particle diameter for the measured critical shear stresses. For uniform, non-cohesive sediments, τ^* can be obtained from the Shields (1936) diagram. Typical values are 0.03, 0.047 and 0.06 (Vanoni 1975). Heterogeneous sediments present additional complications because of hiding and protrusion factors. This issue has been addressed by several researchers, notably Wiberg and Smith (1987), who developed a method to be used for poorly sorted (mixed-size), non-cohesive sediment particles that can also account for variations in particle density.

Cohesive materials. For cohesive streambeds, data obtained with a jet-test device can be used to estimate erosion rates due to hydraulic forces (Hanson 1990, 1991; Hanson and Simon 2001). The rate of erosion ε (m s^{-1}) is assumed to be proportional to the excess shear stress (Foster *et al.* 1977):

$$\varepsilon = k(\tau_0 - \tau_c) = k\tau_e \tag{11.6}$$

where k = erodibility coefficient (m^3 N^{-1} s^{-1}); a is an exponent assumed to be 1.0 and τ_e = excess shear stress (Pa).

An inverse relationship between τ_c and k occurs when soils exhibiting a low τ_c have a high k or when soils having a high τ_c have a low k. Similar trends were observed by Arulanandan *et al.* (1980) during laboratory flume testing of soil samples from cohesive streambed materials obtained across the United States. Based on observations from across the United States, Hanson and Simon (2001) estimated k as a function of τ_c ($r^2 = 0.64$). Here, k is expressed in cm^3 N^{-1} s^{-1}:

$$k = 0.1\tau_c^{-0.5} \tag{11.7}$$

To relate these values to the relative potential for flows to erode cohesive beds and to compare cohesive resistance to the resistance of a non-cohesive particle or aggregate (equivalent diameter), an average boundary shear stress is calculated from eqn. 11.4. The Shields criterion is then invoked to calculate an equivalent particle diameter for the measured critical shear stresses using eqn. 11.5. To calculate erosion rates (ε), values of average boundary or local shear stress are used in conjunction with values of τ_c and k using eqns. 11.6 and 11.7.

Channel width and bank-erosion modelling

Determination of changes in channel width and lateral mobility require that the basic processes of bank retreat and advance be modelled. There has been a significant progress in modelling various *bank erosion processes* (ASCE Task Committee on Hydraulics, Bank Mechanics and Modeling of River Width Adjustment 1998), whereas only recently has progress been made concerning the modelling of *bank advance* (e.g. Parker *et al.* 2010). Quantification of bank retreat involves modelling the two main bank-erosion processes (fluvial erosion and mass failure) and their interaction. A comprehensive review on modelling bank-erosion processes was provided by Rinaldi and Darby (2008).

Fluvial erosion is defined as the removal of bank material by the action of hydraulic forces. The rate of fluvial bank erosion can be quantified using an excess shear stress formula (eqn. 11.6). For granular (non-cohesive) sediments, bank erodibility parameters are modelled based on the same methods that are used to predict the entrainment of bed sediments, albeit with modifications to take into account the effect of the bank angle on the downslope component of the particle weight (Lane 1953) and the case of partly packed and cemented sediments (e.g. Millar and Quick 1993; Millar 2000). For cohesive sediments, critical shear stresses and erodibility coefficients can be obtained with the jet-test device as described earlier. Rates of erosion can then be calculated using the same procedure as described for cohesive beds.

Bank failure occurs when the destabilizing forces, due to gravity, exceed the resisting forces, which are related to the shear strength of the bank materials. The application of stability analyses is common in the bank-erosion literature. The analysis of slide failures is typically performed using a limit equilibrium method (LEM) to compute the factor of safety, defined as the ratio between stabilizing and destabilizing forces. Since the 1980s, specific methods of bank stability analysis have increasingly been developed (e.g. Osman and Thorne 1988; Simon *et al*. 1991; Darby and Thorne 1996), with progressive inclusion of the effects of negative pore-water pressures (Rinaldi and Casagli 1999; Simon *et al*. 1999) and riparian vegetation (Simon and Collison 2002; Pollen *et al*. 2004). Among these analytical methods, a process-based bank-stability model for layered banks that accounts for the effects of pore water and confining pressures was originally developed by Simon and Curini (1998) and Simon *et al*. (1999, 2000). Further enhancement of the Bank-Stability and Toe-Erosion Model (BSTEM) incorporated fluvial erosion and the root reinforcing effects of riparian vegetation (Simon *et al*. 2011). Another example is provided by a simulation approach based on the interaction of the main physical processes (hydrodynamics, groundwater and mass failure), as reported in Rinaldi *et al*. (2008) and Luppi *et al*. (2009).

To quantify the uncertainty of the parameters necessary for the deterministic analyses, probabilistic or reliability analyses can be also performed. In these analyses, a number of parameters related to the highest degree of uncertainty (e.g. cohesion, friction angle) are handled as stochastic variables. This type of probabilistic approach is increasingly being applied to riverbank stability analysis. For example, Parker *et al.* (2008) investigated the effects of variability in bank-material properties on riverbank stability, and Samadi *et al.* (2009) evaluated the extent to which uncertainties in the parameterization of a series of controlling factors influence the reliability of the bank stability results.

Extending the analysis from a single bank profile to lateral mobility, mathematical models offer a rational means of determining channel mobility. However, in practice, modelling lateral change is a challenging problem, since water and sediment discharges vary continuously through time, as well as channel boundary material variability. Examples of reviews on simulation modelling of lateral mobility and on meandering analytical models are those presented by Piégay *et al.* (2005) and Camporeale *et al*. (2007).

11.4 Conclusions

There are many tools available for measuring and analysing channel form, but in lieu of solid quantitative analysis of channel processes, practitioners implementing stream-related projects often rely on generalizations drawn from 1960s-era literature and may apply rules of thumb derived from humid-climate regions to semi-arid regions (Kondolf *et al.* 2013). Some of the reasons offered for the lack of adequate analyses include (i) lack of data, (ii) lack of time, (iii) budget constraints or (iv) personnel limitations. In the present climate of computer modelling, there is the danger that one can easily ignore real data by making broad assumptions and/or creating synthetic data – but collecting field data is at the very core of a usable, reliable model. Collecting basic stream data should be one of the primary goals for anyone planning, designing or implementing projects that are related to the stream corridor.

References

Agnelli, A., Billi, P., Canuti, P. and Rinaldi, M. 1998. *Dinamica Evolutiva Recente dell'Alveo del Fiume Arno*, Monografia CNR-GNDCI, Pubblicazione No. 1739, Pisa: Pacini Editore, 191pp.

Alber, A. and Piégay, H. 2011. Spatial disaggregation and aggregation procedures for characterizing fluvial features at the network-scale: application to the Rhône basin (France). *Geomorphology* 125: 343–360.

Andrews, E.D. 1980. Effective and bankfull discharge of streams in the Yampa River Basin, Colorado and Wyoming. *Journal of Hydrology* 46: 311–330.

Andrews, E.D. and Nankervis, J.M. 1995. Effective discharge and the design of channel maintenance flows for gravel-bed rivers. In: Costa, J.E., Miller, A.J., Potter, K.W. and Wilcock, P.R., eds., *Natural and Anthropogenic Influences in Fluvial Geomorphology*, Geophysical Monograph 89, Washington, DC: American Geophysical Union, pp. 151–164.

Arulanandan, K., Gillogley, E. and Tully, R. 1980. *Development of a Quantitative Method to Predict Critical Shear Stress and Rate of Erosion of Natural Undisturbed Cohesive Soils*, Waterways Experiment Station Technical Report GL-80-5, Davis, CA: University of California, Davis.

ASCE Task Committee on Hydraulics, Bank Mechanics and Modeling of River Width Adjustment 1998. River width adjustment. I. Processes and mechanisms. *Journal of Hydraulic Engineering* 124: 881–902.

Ashmore, P.E. 1991. How do gravel rivers braid? *Canadian Journal of Earth Science* 28: 326–341.

Belletti, B., Dufour, S. and Piégay, H. 2013. Regional variability of aquatic pattern in braided reaches (example of the French Rhône basin). *Hydrobiologia* 712: 25–41.

Bertoldi, W., Zanoni, L. and Tubino, M. 2009. Planform dynamics of braided streams. *Earth Surface Processes and Landforms* 34: 547–557.

Biedenharn, D.S., Combs, P.G., Hill, G.J., Pinkand, C.F. and Pinkston, C.B. 1989. Relationship between channel migration and radius of curvature on the Red River. In: Wang, S.Y., ed., *Sediment Transport Modeling, Proceedings of the International Symposium*, New York: American Society of Civil Engineers, pp. 536–541.

Biedenharn, D.S., Thorne, C.R., Soar, P.J., Hey, R.D. and Watson, C.C. 2001. Effective discharge calculation guide. *International Journal of Sediment Research* 16(4): 445–459.

Blench, T. 1952. Regime theory for self-formed sediment-bearing channels. *Transactions of the American Society of Civil Engineers* 117: 383–408.

Blench, T. 1970. Regime theory design of canals with sand beds. *Journal of Drainage and Irrigation Division, ASCE* 96(IR2): 205–213.

Brasington, J., Rumsby, B.T. and McVey, RA. 2000. Monitoring and modelling morphological change in a braided gravel-bed river using high resolution GPS-based survey. *Earth Surface Processes and Landforms* 25: 973–990.

Brewer P.A. and Passmore, D.G. 2002. Sediment budgeting techniques in gravel bed rivers. In: Jones, S. and Frostick, L., eds., *Sediment Flux to Basins: Causes, Controls and Consequences*, Special Publication 191, London: Geological Society; pp. 97–113.

Brice, J.C. 1964. *Channel Patterns and Terraces of the Loup Rivers in Nebraska*, US Geological Survey Professional Paper 422D, Washington, DC: US Government Printing Office, 41pp.

Brice, J.C. 1975. *Airphoto Interpretation of the Form and Behavior of Alluvial Rivers*, Final Report to the US Army Research Office, Durham, NC.

Brice, J.C. 1984. Planform properties of meandering rivers. In: Elliott, C.M., ed., *River Meandering, Proceedings of Conference on Rivers '83*, New York: ASCE, pp. 1–15.

Brierley, G.J. and Fryirs, K.A. 2005. *Geomorphology and River Management : Applications of the River Style Framework*, Oxford: Blackwell, 398pp.

Camporeale, C., Perona, P., Porporato, A. and Ridolfi, L. 2007. Hierarchy of models for meandering rivers and related morphodynamic processes. *Review of Geophysics*, 45: RG1001.

Carson, M.A. and Kirkby, M.J. 1972. *Hillslope Form and Process*, Cambridge: Cambridge University Press, 475pp.

Castro, J.M. 1996. *Stream Classification in the Pacific Northwest: Methodologies,* Regional Analyses and Applications, PhD Dissertation, Oregon State University, Corvallis, OR.

Castro, J.M. and Jackson, P.L. 2001. Bankfull discharge recurrence intervals and regional hydraulic geometry relationships: patterns in the Pacific Northwest, USA. *Journal of the American Water Resources Association* 37(5): 1249–1262.

Cavalli, M., Tarolli, P., Marchi, L. and Dalla Fontana, G. 2008. The effectiveness of airborne LiDAR data in the recognition of channel-bed morphology. *Catena* 73(3): 249–260.

Chang, H.H. 1988. *Fluvial Processes in River Engineering*, New York: John Wiley & Sons, 432pp.

Church, M.A. 1992. Channel morphology and mypology. In: Callow, P. and Petts, G.E., eds., *The Rivers Handbook*, Oxford: Blackwell, pp. 126–143.

Church, M. and Rice, S.P. 2009. Form and growth of bars in a wandering gravel-bed river. *Earth Surface Processes and Landforms* 10: 1422–1432.

Darby, S.E. and Simon, A. 1999. *Incised River Channels: Processes, Forms, Engineering and Management*, Chichester: John Wiley & Sons.

Darby, S.E. and Thorne, C.R. 1996. Development and testing of river-bank stability analysis. *Journal of Hydraulic Engineering* 122(8): 443–454.

Davis, W.M. 1909. *Geographical Essays*, Boston: Ginn (republished 1954, Dover, New York).

Downs, P.W. and Thorne, C.R. 1996. A geomorphological justification of river channel reconnaissance surveys. *Transactions of the Institute of British Geographers* NS21: 455–468.

Egozi, R. and Ashmore, P. 2008. Defining and measuring braiding intensity. *Earth Surface Processes and Landforms* 33: 2121–2138.

Elliott, J.G., Gellis, A.C. and Aby, S.B. 1999. Evolution of arroyos: incised channels of the southwestern United States. In: Darby, S.E. and Simon, A., eds., *Incised Channels: Processes, Forms, Engineering and Management*, Chichester: John Wiley & Sons, pp. 153–186.

Foster, G.R., Meyer, L.D. and Onstad, C.A. 1977. An erosion equation derived from basic erosion principles. *Transactions of the American Society of Agricultural Engineers* 20: 678–682.

Fuller, I.C., Reid, H.E. and Brierley, G.J. 2013. Methods in geomorphology: investigating river channel form. In: Shroder, J.F., ed., *Treatise on Geomorphology*, vol. 14, San Diego: Academic Press, pp. 73–91.

Geerling, G.W., Vreeken-Buijs, M.J., Jesse, B., Ragas, A.M.J. and Smits, A.J.M. 2009. Mapping river floodplain ecotopes by segmentation of spectral (CASI) and structural (LiDAR) remote sensing data. *River Research and Applications* 25: 795–813.

Gilbert, G.K. 1914. *The Transportation of Debris by Running Water*, US Geological Survey Professional Paper 86, Washington, DC: US Government Printing Office.

Graf, W.L. 1977. The rate law in fluvial geomorphology. *American Journal of Science* 277: 178–191.

Gregory K.J. and Walling, D.E. 1973. *Drainage Basin Form and Process: a Geomorphological Approach*, London: Edward Arnold.

Gurnell, AM. 1997. Channel change of the River Dee meanders 1946–1992, from the analysis of air photographs. *Regulated Rivers – Research and Management* 13: 13–26.

Gurnell, A., Surian, N. and Zanoni, L. 2009. Multi-thread river channels: a perspective on changing European alpine river systems. *Aquatic Sciences* 71: 253–265.

Halwas, K.L. and Church, M. 2002. Channel units in small, high gradient streams on Vancouver Island, British Columbia. *Geomorphology* 43: 243–257.

Hanson, G.J. 1990. Surface erodibility of earthen channels at high stresses. Part II – Developing an in-situ testing device. *Transactions of the ASAE* 33(1): 132–137.

Hanson, G.J. 1991. Development of a jet index to characterize erosion resistance of soils in earthen spillways. *Transactions of the ASAE*, 36(5): 1349–1352.

Hanson, G.J. and Simon, A. 2001. Erodibility of cohesive streambeds in the loess area of the midwestern USA. *Hydrological Processes* 15(1): 23–38.

Harrelson, C.C., Rawlins, C.L. and Potyondy, J.P. 1994. *Stream Channel Reference Sites: an Illustrated Guide to Field Technique*, General Technical Report RM-245, Fort Collins, CO: US Department of Agriculture, Forest Service, 61pp.

Hassan, M.A., Egozi, R. and Parker, G. 2006. Experiments on the effect of hydrograph characteristics on vertical grain sorting in gravel bed rivers. *Water Resources Research*, 42: W09408.

Hey, R.D. and Thorne, C.R. 1986. Stable channels with mobile gravel beds. *Journal of Hydraulic Engineering* 112(8): 671–689.

Hicks, D.M., Duncan, M.J., Lane, S.N., Tal, M. and Westaway, R. 2008. Contemporary morphological change in braided gravel-bed rivers: new developments from field and laboratory studies, with particular reference to the influence of riparian vegetation. In: Habersack, H., Piégay, H. and Rinaldi, M., eds., *Gravel-Bed Rivers 6 – From Process Understanding to River Restoration*, Series Developments in Earth Surface Processes, vol. 11, Amsterdam: Elsevier, pp. 557–584.

Hooke, J.M. 2007. Spatial variability, mechanisms and propagation of change in active meandering river. *Geomorphology* 84: 277–296.

Hughes, M.L., McDowell, P.F. and Marcus, WA. 2006. Accuracy assessment of georectified aerial photographs: implications for measuring lateral channel movement in a GIS. *Geomorphology* 74: 1–16.

Inglis, C.C. 1949. The Behaviour and Control of Rivers and Canals, *Research Publication 13*, Poona: Central Waterpower Irrigation and Navigation Research Station.

Jacobson, R.B. 1995. Spatial controls on patterns of land-use induced stream disturbance at the drainage-basin scale – an example from gravel-bed streams of the Ozark Plateaus, Missouri. In: Costa, J.E., Miller, A.J., Potter, K.W. and Wilcock, P.R., eds., *Natural and Anthropogenic Influences in Fluvial Geomorphology*, Geophysical Monograph 89 Washington, DC: American Geophysical Union, pp. 219–239.

Knighton, D. 1982. Longitudinal changes in size and shape of stream bed material: evidence of variable transport conditions. *Catena* 9: 25–34.

Knighton, D. 1998. *Fluvial Forms and Processes*, New York: John Wiley & Sons, 383pp.

Kondolf, G.M., Podolak, K. and Grantham, T.E. 2013. Restoring Mediterranean-climate rivers. *Hydrobiologia* 719: 527–545.

Lacey, G. 1930. Stable channels in alluvium. *Proceedings of the Institution of Civil Engineers (London)* 229: 259–384.

Lacey, G. 1958. Flow in alluvial channels with sand mobile beds. *Proceedings of the Institution of Civil Engineers (London)* 9 (discussion 11).

Lane, E.W. 1953. Progress report on studies on the design of stable channels of the Bureau of Reclamation. *Proceedings of the American Society of Civil Engineers* 79: 246–261.

Lane, E.W. 1955. Design of stable alluvial channels. *Transactions, American Society of Civil Engineers* 120(2776): 1234–1260.

Lane, S.N. and Chandler, J.H. 2003. Editorial. The generation of high quality topographic data for hydrology and geomorphology: new data sources, new applications and new problems. *Earth Surface Processes and Landforms* 28(3): 229–230.

Langendoen, E. J. 2000. *CONCEPTS – CONservational Channel Evolution and Pollutant Transport System*, Research Report No. 16, Oxford, MS: US Department of Agriculture, Agricultural Research Service, National Sedimentation Laboratory.

Laronne, J.B., Outhet, D.N., Carling, P.A. and McCabe, T.J. 1994. Scour chain employment in gravel bed rivers. *Catena* 22: 299–306.

Lawler, D.M. 1993. The measurement of river bank erosion and lateral channel change: a review. *Earth Surface Processes and Landforms* 18: 777–821.

Lawler, D.M. 2005. Defining the moment of erosion: the principle of thermal consonance timing. *Earth Surface Processes and Landforms* 30(13): 1597–1615.

Lawler, D.M. 2007. Advances in the continuous monitoring of erosion and deposition dynamics: developments and applications of the new PEEP-3T system. *Geomorphology* 93: 17–39.

Lawler, D.M., Bull, L. and Harris, N.M. 1997. Bank erosion events and processes in the Upper Severn. *Hydrological and Earth System Science*, 1: 523–534.

Lejot, J., Delacourt, C., Piégay, H., Fournier, T., Trémélo, M.-L. and Allemand, P. 2007. Very high spatial resolution imagery for channel bathymetry and topography from an unmanned mapping controlled platform. *Earth Surface Processes and Landforms* 32(11): 1705–1725.

Leopold, L.B. 1994. *A View of the River*. Cambridge, MA: Harvard University Press.

Leopold, L.B. and Maddock, T. Jr. 1953. *The Hydraulic Geometry of Stream Channels and Some Physiographic Implications*, US Geological Survey Professional Paper 252, Washington, DC: US Government Printing Office.

Leopold, L.B. and Wolman, M.G. 1957. *River Channel Patterns – Braided, Meandering and Straight*, US Geological Survey Professional Paper 282B, Washington, DC: US Government Printing Office, pp. 39–85.

Leopold, L.B., Wolman, M.G. and Miller, J.P. 1964. *Fluvial Processes in Geomorphology*, New York: W.H. Freeman, 522pp.

Lichti, D.D., Gordon, S.J. and Stewart, M.P. 2002. Ground-based laser scanners: operation, systems and applications. *Geomatica* 56(1): 21–33.

Liébault, F., Lallias-Tacon, S., Cassel, M. and Talaska, N. 2012. Long profile responses of alpine braided rivers in SE France. *River Research and Applications* 29(10): 1253–1266.

Lisle, T.E. and Eads, R. 1991. *Methods to Measure Sedimentation of Spawning Gravels*, Research Note PSW-411, Berkeley, CA: US Department of Sgriculture Forest Service, Pacific Southwest Research Station.

Little, W.C., Thorne, C.R. and Murphy, J.B. 1982. Mass bank failure analysis of selected Yazoo Basin streams. *Transactions of the American Society of Agricultural Engineeers* 25: 1321–1328.

Lohnes, R.L. 1991. A method for estimating land loss associated with stream channel degradation. *Engineering Geology* 31: 115–130.

Lohnes, R.A. and Handy, R.L. 1968. Slope angles in friable loess. *Journal of Geology* 76(3): 247–258.

Luppi, L., Rinaldi, M., Teruggi, L.B., Darby, S.E. and Nardi, L. 2009. Monitoring and numerical modelling of riverbank erosion processes: a case study along the Cecina River (Central Italy). *Earth Surface Processes and Landforms* 34 (4): 530–546.

Lutenegger, J.A. and Hallberg, B.R. 1981. Borehole shear test in geotechnical investigations. In: Yong, R.N. and Townsend, F.C., eds., *Laboratory Shear Strength of Soil*, ASTM Special Publication STP 740, Philadelphia, PA: American Society for Testing and Materials, pp. 566–578.

Mackin, J.H. 1948. Concept of a graded river. *Geological Society of America Bulletin* 59: 463–511.

Malavoi, J.R. and Bravard, J.P. 2010. *Elements d'Hydromorphologie Fluviale*, Baume-les-Dames: ONEMA, 224pp.

Marcus, W.A. and Fonstad, M.A. 2008. Optical remote mapping of rivers at sub-meter resolutions and watershed extents. *Earth Surface Processes and Landforms* 33(1): 4–24.

McLean, D.G. and Church, M. 1999. Sediment transport along lower Fraser River: 2. Estimates based on the long-term gravel budget. *Water Resources Research* 35(8): 2549–2559.

Milan, D.J., Heritage G.L. and Hetherington, D. 2007. Application of a 3D laser scanner in the assessment of erosion and deposition volumes and channel change in a proglacial river. *Earth Surface Processes and Landforms* 32(11): 1657–1674.

Millar, R.G. 2000. Influence of bank vegetation on alluvial channel patterns. *Water Resources Research* 36(4): 1109–1118.

Millar, R.G. and Quick, M.C. 1993. Effect of bank stability on geometry of gravel rivers. *Journal of Hydraulic Engineering* 119: 1343–1363.

Miller, J.R. and Ritter, J.B. 1996. An examination of the Rosgen classification of natural rivers. Discussion. *Catena* 27: 295–299.

Molinas, A. 1989. *User's Manual for BRI-STARS*, National Cooperative Highway Research Program, Project No. HR-15-11, Fort Collins, CO Hydrau-Tech Engineering and Software, 156pp.

Montgomery, D.R. and Buffington, J.M. 1997. Channel-reach morphology in mountain drainage basins. *Geological Society of America Bulletin* 109(5): 596–611.

Morisawa, M. 1968. *Streams: Their Dynamics and Morphology*, New York: McGraw-Hill.

Mount, N.J., Louis, J., Teeuw, R.M., Zukowski, P.M. and Stott, T. 2003. Estimation of error in bankfull width comparisons from temporally sequenced raw and corrected aerial photographs. *Geomorphology* 56: 65–77.

Murphey, J.B. and Grissinger, E.H. 1985. Channel cross-section changes in Mississippi's Goodwin Creek. *Journal of Soil and Water Conservation* 40(1): 148–153.

O'Neal, M.A. and Pizzuto, J.E. 2011. The rates and spatial patterns of annual riverbank erosion revealed through terrestrial laser-scanner surveys of the South River, Virginia. *Earth Surface Processes and Landforms* 36(5): 695–701.

Osman, A.M. and Thorne, C.R. 1988. Riverbank stability analysis. I. Theory. *Journal of Hydraulic Engineering* 114(2): 134–150.

Parker, C., Simon, A. and Thorne, C.R. 2008. The effects of variability in bank material properties on riverbank stability: Goodwin Creek, Mississippi. *Geomorphology* 101: 533–543.

Parker, G., Shimizu, Y., Wilkerson, G.V., Eke, E.C., Abad, J.D., Lauer, J.W., Paola, C., Dietrich, W.E. and Voller, V.R. 2010. A new framework for modeling the migration of meandering rivers. *Earth Surface Processes and Landforms* 36: 70–86.

Pellegrini, L., Maraga, F., Turitto, O., Audisio, C. and Duci, G. 2008. Evoluzione morfologica di alvei fluviali mobili nel settore occidentale del bacino padano. *Il Quaternario* 21(1B): 251–266.

Petts, G.E., Müller, H. and Roux, A.L. 1989. *Historical Change of Large Alluvial Rivers: Western Europe*, Chichester: John Wiley & Sons.

Piégay, H., Darby, S.E., Mosselman, E. and Surian, N. 2005. A review of techniques available for delimiting the erodible river corridor: a sustainable approach to managing bank erosion. *River Research and Applications* 21: 773–789.

Pollen, N., Simon, A. and Collison, A. 2004. Advances in assessing the mechanical and hydrologic effects of riparian vegetation on streambank stability. In: Bennett, S.J. and Simon, A., eds., *Riparian Vegetation and Fluvial Geomorphology*, Washington, DC: American Geophysical Union, pp. 125–140.

Pyle, C.J., Chandler, J.H. and Richards, K.S 1997. Digital photogrammetric monitoring of riverbank erosion. *Photogrammetric Record* 15: 753–764.

Rice, S.P. 1998. Which tributaries disrupt downstream fining along gravel-bed rivers? *Geomorphology* 22: 39–56.

Richards, K. 1982. *Rivers: Form and Process in Alluvial Channels*, London: Methuen, 358pp.

Rinaldi, M. 2003. Recent channel adjustments in alluvial rivers of Tuscany, Central Italy. *Earth Surface Processes and Landforms* 28(6): 587–608.

Rinaldi, M. 2008. Schede di rilevamento geomorfologico di alvei fluviali. *Il Quaternario* 21(1B): 353–366.

Rinaldi, M. and Casagli, N. 1999. Stability of streambanks formed in partially saturated soils and the effects of negative pore water pressures: the Sieve River (Italy). *Geomorphology* 26: 253–277.

Rinaldi, M. and Darby, S.E. 2008. Modelling river-bank-erosion processes and mass failure mechanisms: progress towards fully coupled simulations. In: Habersack, H., Piégay, H. and Rinaldi, M., eds., *Gravel-Bed Rivers 6 – From Process Understanding to River Restoration*, Series Developments in Earth Surface Processes, vol. 11, Amsterdam: Elsevier, pp. 213–239.

Rinaldi, M. and Simon, A. 1998. Bed-level adjustments in the Arno River, central Italy. *Geomorphology* 22: 57–71.

Rinaldi, M., Simon, A. and Billi, P. 1997. Disturbance and adjustment of the Arno River, Central Italy. II. Quantitative analysis of the last 150 years. In: Wang, S.S.Y., Langendoen, E.J. and Shields F.D. Jr, eds., *Management of Landscapes Disturbed by Channel Incision, Stabilization, Rehabilitation, Restoration*, Oxford, MS: Center for the Computational Hydroscience and Engineering, University of Mississippi, pp. 601–606.

Rinaldi, M., Mengoni, B., Luppi, L., Darby, S.E. and Mosselman, E. 2008a. Numerical simulation of hydrodynamics and bank erosion in a river bend. *Water Resources Research* 44: W09429.

Rinaldi, M., Teruggi, L.B., Simoncini, C. and Nardi, L. 2008b. Dinamica recente ed attuale di alvei fluviali: alcuni casi di studio dell'Appennino Settentrionale. *Il Quaternario* 21(1B): 291–302.

Rosgen, D.L. 1996. *Applied River Morphology*. Pagosa Springs, CO: Wildland Hydrology, 390pp.

Rust, B.R. 1978. A classification of alluvial systems. In: Miall, A.D., ed., Fluvial Sedimentology, *Memoir 5*, Calgary: Canadian Society of Petroleum Geologists, pp. 187–198.

Samadi, A., Amiri-Tokaldany, E. and Darby, S.E. 2009. Identifying the effects of parameter uncertainty on the reliability of riverbank stability modelling. *Geomorphology* 106: 219–230.

Schumm, S.A. 1977. *The Fluvial System*, New York: John Wiley & Sons, 338pp.

Schumm, S.A. 1981. Evolution and response of the fluvial system, sedimentologic implications. In: Ethridge, F.G. and Flores, R.M., eds., *Recent and Ancient Nonmarine Depositional Environments: Models for Exploration*, Special Publication 31, Tulsa, OK: Society of Economic Paleontologists and Mineralogists, pp. 19–29.

Schumm, S.A. 1985. Patterns of alluvial rivers. *Annual Reviews of Earth and Planetary Science* 13: 5–27.

Schumm, S.A. and Lichty, R.W. 1965. Time, space and causality in geomorphology. *American Journal of Science* 263: 110–119.

Schumm, S.A., Harvey, M.D. and Watson, C.C. 1984. *Incised Channels: Morphology,* Dynamics and Control, Littleton, CO: Water Resources Publications, 200pp.

Shields, A. 1936. Anwendung der Ähnlichkeitsmechanik und Turbulenzforschung auf die Geschiebebewegung. *Mitteil. Preuss. Versuchsanst. Wasser*, Erd, Schiffsbau, Berlin, Nr. 26.

Shields, F.D. Jr, Copeland, R.R., Klingeman, P.C., Doyle, M.W. and Simon, A. 2003. Design for stream restoration. *Journal of Hydraulic Engineering* 129: 575–584.

Simon, A. 1989a. A model of channel response in disturbed alluvial channels. *Earth Surface Processes and Landforms* 14: 11–26.

Simon, A. 1989b. The discharge of sediment in channelized alluvial streams. *Water Resources Bulletin* 25(6): 1177–1188.

Simon, A. 1992. Energy, time and channel evolution in catastrophically disturbed fluvial systems. *Geomorphology* 5: 345–372.

Simon, A. and Collison, A.J.C. 2002. Quantifying the mechanical and hydrologic effects of riparian vegetation on streambank stability. *Earth Surface Processes and Landforms* 27: 527–546.

Simon, A. and Curini, A. 1998. Pore pressure and bank stability: the influence of matric suction. In: Abt, S.R., Young-Pezeshk, J. and Watson, C.C., eds., *Water Resources Engineering '98*, New York: American Society of Civil Engineers, pp. 358–363.

Simon, A. and Downs, P.W. 1995. An interdisciplinary approach to evaluation of potential instability in alluvial channels. *Geomorphology* 12(3): 215–232.

Simon, A. and Hupp, C.R. 1986. Channel evolution in modified Tennessee channels. In: *Proceedings of the Fourth Federal Interagency Sedimentation Conference, Las Vegas, March 24–27 1986*, vol.2, pp. 2: 5-71–5-82.

Simon, A. and Hupp, C.R. 1992. *Geomorphic and Vegetative Recovery Processes Along Modified Stream Channels of West Tennessee*, US Geological Survey Open-File Report 91-502, Nashville, TN: US Geological Survey, 142pp.

Simon, A. and Rinaldi, M. 2000. Channel instability in the loess area of the Midwestern United States. *Journal of the American Water Resources Association* 36(1): 133–150.

Simon, A. and Rinaldi, M. 2006. Disturbance, stream incision and channel evolution: the roles of excess transport capacity and boundary materials in controlling channel response. *Geomorphology* 79: 361–383.

Simon, A., Wolfe, W.J. and Molinas, A. 1991. Mass wasting algorithms in an alluvial channel model, In: *Proceedings of the 5th Federal Interagency Sedimentation Conference, Las Vegas, Nevada*, vol. 2, pp. 8-22–8-29.

Simon, A., Curini, A., Darby, S.E. and Langendoen, E. 1999. Streambank mechanics and the role of bank and near-bank processes in incised channels. In: Darby, S.E. and Simon, A., eds., *Incised Channels: Processes,* Forms, Engineering and Management, Chichester: John Wiley & Sons, pp. 123–152.

Simon, A., Curini, A., Darby, S.E. and Langendoen, E.J. 2000. Bank and near-bank processes in an incised channel. *Geomorphology* 35: 193–217.

Simon, A., Dickerson, W. and Heins, A. 2004. Suspended-sediment transport rates at the 1.5-year recurrence interval for ecoregions of the United States: transport conditions at the bankfull and effective discharge? *Geomorphology* 58: 243–262.

Simon, A., Pollen-Bankhead, N. and Thomas, R.E. 2011. Development and application of a deterministic bank stability and toe erosion model for stream restoration. In: Simon, A., Bennett, S.J. and Castro, J.M., eds., *Stream Restoration in Dynamic Fluvial Systems: Scientific Approaches, Analyses and Tools*, Geophysical Monograph Series 194, Washington, DC: American Geophysical Union, pp. 453–474.

Simons, D.B. and Albertson, M.L. 1960. Uniform water conveyance channels in alluvial material. *Journal of the Hydraulics Division, American Society of Civil Engineers* 86(HY5): 33–71.

Skempton, A.W. 1953. Soil mechanics in relation to geology. *Proceedings, Yorkshire Geological Society* 29: 33–62.

Soar P.J. and Thorne, C.R. 2012. Design discharge for river restoration. In: Simon, A., Bennett, S.J. and Castro, J.M., eds., *Stream Restoration in Dynamic Fluvial Systems: Scientific Approaches, Analyses and Tools*, Geophysical Monograph Series 194, Washington, DC: American Geophysical Union, pp. 123–149.

Stevens, H.H. Jr, and Yang, C.T. 1989. *Summary and Use of Selected Fluvial Sediment-Discharge Formulas*, US Geological Survey Water Resources Investigations Report 89-4026, Denver, CO: US Geological Survey, 62pp.

Stott, T.A. 2005. Natural recovery from accelerated forest ditch and stream bank erosion five years after harvesting of plantation forest on Plynlimon, mid-Wales. *Earth Surface Processes and Landforms* 30: 349–357.

Surian, N. and Rinaldi, M. 2003. Morphological response to river engineering and management in alluvial channels in Italy. *Geomorphology* 50: 307–326.

Surian, N., Rinaldi, M. and Pellegrini, L. 2009a. *Linee Guida per l'Analisi Geomorfologica degli Alvei Fluviali e delle Loro Tendenze Evolutive*, Padua: Casa Editrice CLEUP, 78pp.

Surian, N., Rinaldi, M., Pellegrini, L., Audisio, C., Maraga, F., Teruggi, L.B., Turitto, O. and Ziliani L. 2009b. Channel adjustments in Italian rivers: evolutionary trends, causes and management implications. In: James, L.A., Rathburn, S.L. and Whittecar, G.R., eds., *Management and Restoration of Fluvial Systems with Broad Historical Changes and Human Impacts*, Special Paper 451, Boulder, CO: Geological Society of America, pp. 83–95.

Swanson, B.J., Meyer, G.A. and Coonrod, J.E. 2011. Historical channel narrowing along the Rio Grande near Albuquerque, New Mexico in response to peak discharge reductions and engineering: magnitude and uncertainty of change from air photo measurements. *Earth Surface Processes and Landforms* 36: 885–900.

Thorne, C.R. 1997. Channel types and morphological classification. In: Thorne, C.R., Hey, R.D. and Newson, M.D., eds., *Applied Fluvial Geomorphology for River Engineering and Management*, Chichester: John Wiley & Sons, pp. 175–222.

Thorne, C.R. 1998. *Stream Reconnaissance Handbook*, Chichester: John Wiley & Sons, 133pp.

Thorne, C.R. and Osman, M.A. 1988. Riverbank stability analysis. Part II. Application. *Journal of the Hydraulic Engineering* 114(2): 151–171.

Thorne, C.R., Murphey, J.B. and Little, W.C. 1981. *Stream Channel Stability, Appendix D, Bank Stability and Bank Material Properties in the Bluffline Streams of Northwest Mississippi*, Oxford, MS, US Department of Agriculture Sedimentation Laboratory, 257pp.

Thorne, C.R., Russell, A.P.G. and Alam, M.-K. 1993. Planform pattern and channel evolution of the Brahmaputra River, Bangladesh. In: Best, J.L. and Bristow, C.S., eds., *Braided Rivers*, Geological Society Special Publication No. 75, London: Geological Society, pp. 257–276.

Thorne, C.R., Allen, R.G. and Simon, A. 1996. Geomorphological river channel reconnaissance for river analysis, engineering and management. *Transactions of the Institute of British Geographers* NS21: 469–483.

Turowski, J.M., Hovius, N., Wilson, A. and Horng, M.J. 2008. Hydraulic geometry, river sediment and the definition of bedrock channels. *Geomorphology* 99: 26–38.

Vanoni, V.A. 1975, *Sedimentation Engineering*, ASCE Manuals and Reports on Engineering Practice No. 54, Reston, VA: American Society of Civil Engineers, 745pp.

Welber, M., Bertoldi, W. and Tubino, M. 2012. The response of braided planform configuration to flow variations, bed reworking and vegetation: the case of the Tagliamento River, Italy. *Earth Surface Processes and Landforms* 37: 572–582.

Westaway, R.M., Lane, S.N. and Hicks, D.M. 2001. Remote sensing of clear-water, shallow, gravel-bed rivers using digital photogrammetry. *Photogrammetric Engineering and Remote Sensing* 67(11): 1271–1281.

Wheaton, J.M., Brasington, J., Darby, S.E. and Sear, D.A. 2010. Accounting for uncertainty in DEMs from repeat topographic surveys: improved sediment budgets. *Earth Surface Processes and Landforms* 35: 136–156.

Wiberg, P.L. and Smith J.D. 1987. Calculations of the critical shear stress for motion of uniform and heterogeneous sediments. *Water Resources Research* 23: 1471–1480.

Williams, G.P. 1978a. Bankfull discharge of rivers. *Water Resources Research* 14: 1141–1154.

Williams, G.P. 1978b. *Hydraulic Geometry of River Cross-sections – Theory of Minimum Variance*, US Geological Survey Professional Paper 1029, Washington, DC: US Government Printing Office.

Williams, G.P. and M.G. Wolman 1984. *Downstream Effects of Dams on Alluvial Rivers*, US Geological Survey Professional Paper 1286, Washington, DC: US Government Printing Office, 83pp.

Williams, R.P. 1979. *Sediment Discharge in the Santa Clara River Basin, Ventura and Los Angeles Counties, California*, Water-Resources Investigations 79-78, Menlo Park, CA: US Geological Survey.

Wilson, K.V. Jr and Turnipseed, D.P. 1993. Channel-bed and Channel-bank Stability of Standing Pine Creek Tributary at State Highway 488 at Free Trade, Leake County, Mississippi. US Geological Survey Open-File Report 93-37, Jackson, MS: US Geological Survey, 20pp.

Wilson, K.V. Jr and Turnipseed, D.P. 1994. *Geomorphic Response to Channel Modifications of Skuna River at the State Highway 9 Crossing at Bruce, Calhoun County, Mississippi*, US Geological Survey Water-Resources Investigations Report 94-4000, Jackson, MS: US Geological Survey, 43pp.

Winterbottom, S.J. 2000. Medium and short-term channel planform changes on the Rivers Tay and Tummel, Scotland. *Geomorphology* 34: 195–208.

Wolman, M.G. and Gerson, R. 1978. Relative scales of time and effectiveness of climate in watershed geomorphology. *Earth Surface Processes* 3: 189–208.

Wolman, M.G. and Miller, J.P. 1960. Magnitude and frequency of forces in geomorphic processes. *Journal of Geology* 68: 58–74.

Yevjevich, V. 1972. *Probability and Statistics in Hydrology*, Fort Collins, CO: Water Resources Publications.

CHAPTER 12

Flow measurement and characterization

Peter J. Whiting
Case Western Reserve University, Cleveland, OH, USA

12.1 Introduction

A basic tool of the geomorphologist and hydrologist is the measurement of streamflow, to determine stream discharge, to estimate the flow resistance and boundary shear stress and to characterize the turbulence and other flow attributes. There are many examples of the need for such information. The discharge at upstream locations is usually needed to anticipate downstream flood levels and to decide whether to open or close dam spillways. These decisions can affect public safety and have significant societal, environmental and monetary costs. Water allocation and administration require accurate discharge measurement, as does the construction of a basic water budget, for example, to determine the effect of silvicultural activities on the volume and timing of runoff. Sediment discharge is often related to flow discharge, hence sediment budgets often depend upon flow discharge estimation. The characterization of local flow velocity or turbulence is often needed in process-based studies. Moreover, discharge is often determined from local measurements of flow velocity, hence the ability to characterize the magnitude and direction of flow through the water column is critical. Velocity profiles can be used to estimate the magnitude of the local boundary shear stress from the law of the wall. Very detailed profiles can reveal the magnitude of the resistance to flow associated with bedforms and banks. Faithful characterization of the flow field is one test of the ability of two- and three-dimensional models to predict flow patterns. Fisheries biologists often estimate the potential utilization of habitat based upon the availability of areas with characteristic ranges of velocity. Efforts to understand sediment transport processes better often link sediment particles to boundary shear stress and turbulence.

A variety of equipment and techniques are available to measure flow and discharge in the field and the range of technologies is expanding. These new technologies can improve the accuracy and precision of measurements, speed the collection of data and/or provide new information on the flow. This chapter describes and compares available methods to measure velocity and flow in some detail and explores issues that should be considered in selecting a method for measuring flow in the context of fluvial geomorphology. More comprehensive treatments of the subject, at least for standard methods, can be found in *Streamflow Measurement* (Herschy 1985), *Water Measurement Manual* (Bureau of Reclamation 1984) and *Measurement of Liquid Flow in Open Channels* (ISO 1983), among other books. This chapter will be most useful in helping scientists and decision-makers select the most appropriate methods for their specific problems. Details about implementing specific methods are presented in the references cited in the text.

12.2 Velocity measurement

In this section, the techniques and equipment for measuring flow velocity are described, the principles underlying the approaches explained and the accuracy and appropriateness of the approaches discussed.

An example of the measurement of velocity comes from my work at Solfatara Creek, Wyoming (USA), where I sought to understand the downstream and cross-stream accelerations of flow induced by rapid shoaling associated with a mid-channel bar (Whiting and Dietrich 1991). A key part of examining the magnitude of the accelerations was the characterization of the flow field at many points in the water column, at multiple points across the channel and at multiple closely spaced cross-sections. Both the downstream and cross-stream components of velocity were measured at each point. Multiple points had to be measured in the water column for several reasons. One reason was to define the lateral and downstream fluxes that varied with height in the water column. Near-bed measurements were also needed to provide an estimate of local boundary shear stress at each vertical across each section. Finally, the shoaling was dramatic enough that fine spatial resolution was required to investigate the phenomena.

Floats

Water velocity can be estimated from floats (Dunne and Leopold 1978; Herschy 1985). In situations where another technique is inappropriate or too hazardous or the proper equipment is unavailable, the downstream displacement over time of

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

buoyant objects such as sticks or oranges can be measured. It is recommended that travel time be at least 20 s. Floats provide a quick estimate of surface velocity, but the accuracy of such an estimate of velocity is less than that of other methods, probably no better than ±10–20%. The average velocity at a location is often estimated by multiplying the surface velocity by 0.85 (0.8 for rocky channel bottoms and 0.9 for muddy bottoms). Christensen (1994) described a variation of the float technique in which an orange is held at the bottom and released, and the time to reach the surface and then float a characteristic distance is measured. The only equipment required for these float methods is a watch and a tape measure (and the float).

An alternative use of floats is to determine gross patterns of flow or the location of eddies. Thus floats can be helpful in the selection of cross-sections for measurement or in the selection of the measurement approach.

Mechanical current meters

Mechanical current meters have been the most commonly used equipment for measuring flow velocity for decades, although their use is declining. Mechanical current meters measure flow velocity from the rotation of a vertical-axis bucket-wheel with cups or a horizontal-axis impeller. In standard operation, the rotational speed of the cups or impeller is derived from the number of revolutions per unit time as measured optically, magnetically or electrically. Determining the number of rotations by counting 'clicks', as a circuit is completed, is rarely done any longer.

The vertical-axis meters include the larger Price AA and smaller Price mini-meter (or 'pygmy' meter) (Fig. 12.1a and b). The diameter of the bucket wheel of the meters is 13 and 5 cm, respectively. The bucket wheels should be constructed of metal rather than plastic (Jarrett 1992). The Price AA can be used in water as shallow as 0.15 m and to measure flow velocities from 0.06 to almost 8 m s^{-1}. The smaller mini-meter was designed for use in shallower and/or slower flows: water depths from 0.08 to 0.45 m and water velocities from 0.02 to 0.9 m s^{-1}. In principle, there is no problem using the mini-meter in deeper flow if flow velocity is in the appropriate range. The Price and mini-meters are often deployed with a vane serving to orient the meter in the flow. The calibration of these devices for flow approaching the meter at an angle has been determined (Fulford *et al.* 1994). The accuracy of the magnitude of the flow velocity of Price AA and mini-meters is about 0.5% (Fulford *et al.* 1994). While Price meters have been used to measure turbulence, the frequency response of both meters is less than 1 Hz, hence they are not capable of quantifying the higher frequency part of the turbulent spectrum.

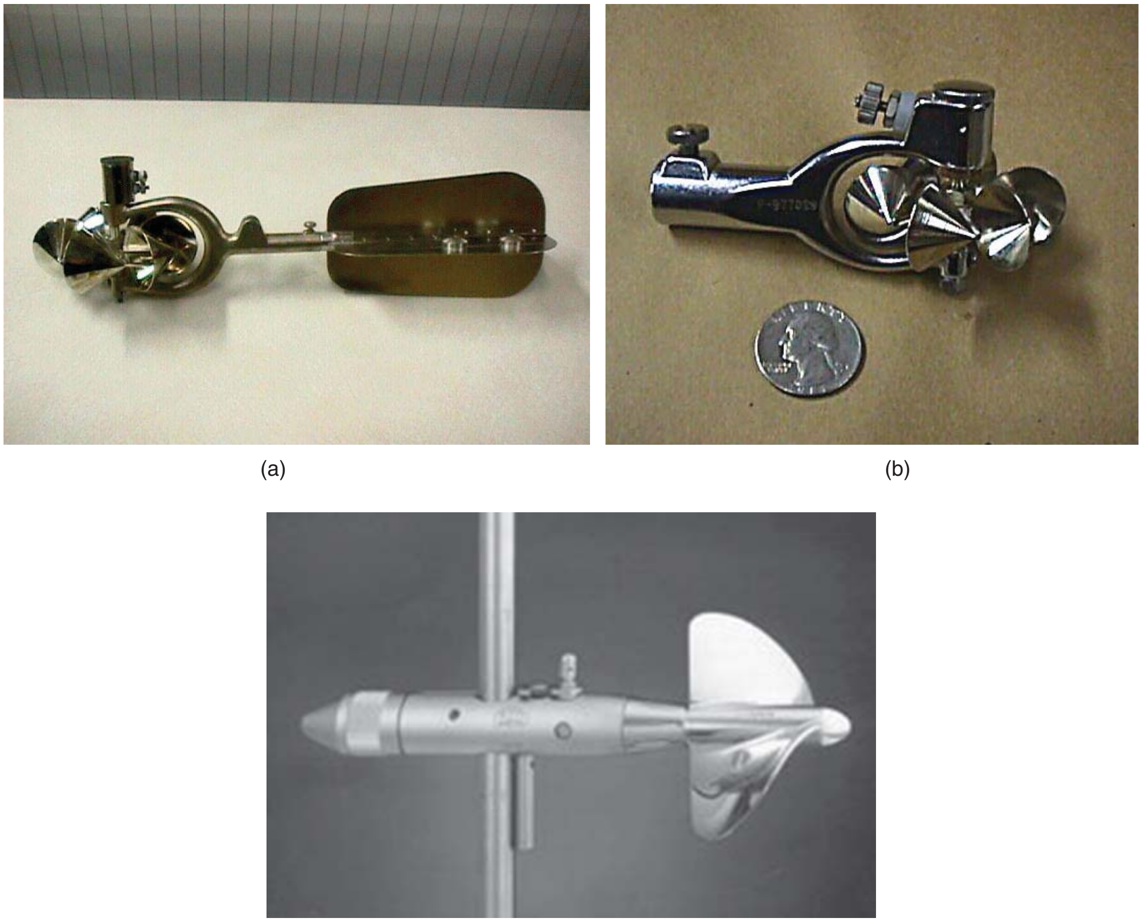
The horizontal-axis meters include the Ott meter (Fig. 12.1c) and Smith meter (Smith 1978), among others. These meters have a screw-type impeller that is typically 5–8 cm across and the meters are capable of measuring flow from 0.05 to 8 m s^{-1}. Other miniature impeller current meters appropriate for the field, nonetheless fragile, can be as small as 1.2 cm in diameter. They are capable of measuring velocities from 0.03 to 3 m s^{-1}. The calibration of these devices for flow approaching the meter at other than parallel to the spindle of the impeller varies by model, if it is known at all. The accuracy of flow velocity varies widely by model and manufacturer; the range is about 0.75–2.0% (Fulford *et al.* 1994).

Comparing the two types of mechanical current meters, the vertical-axis meters can be used at lower flow velocity (except perhaps the fragile miniature screw-type meters), but disturb the flow more and are more prone to becoming tangled by debris or growing vegetation than horizontal-axis meters (Fulford *et al.* 1994). The uncertainties associated with vertical- and horizontal-axis meters are similar but in general the accuracy of the vertical-axis meters is higher (Fulford *et al.* 1994).

The mechanical current meters, with the exception of the miniature meters, are very robust in the field. Even if damaged, many repairs are possible in the field. The maintenance requirements are modest, particularly for the vertical-axis meters. The equipment should be cleaned daily after use. Prior to use, the spin of the rotor should be checked. It should take over 30 s for the rotor to stop spinning after it is spun by hand in air (Rantz *et al.* 1982).

Current meters (mechanical and other types) are deployed from wading rods, cables, bridges or other structures and boats. Top-set wading rods allow the hydrographer to stand in the flow and to re-position the meter in the water column without reattaching the meter to the support and without removing the wading rod from the water or alternatively getting their hands wet (Fig. 12.2a). When the water depth or velocity is too large for wading, meters can be suspended from a cableway (Fig. 12.2b) or bridge or boat (Fig. 12.2c). A weight can be used to submerge the suspended current meter. A weight is attached to the base of a cable to maintain its position in the streaming flow deflecting the cable downstream. Various equipment for use in suspending current meters (reels, weights, cables, bridge boards, etc.) is described in Buchanan and Somers (1969) and Herschy (1985). If flow and depth permit, measurement by wading is often preferred because of the greater control that the hydrographer can employ in the holding and placement of the current meter and wading rod. For very detailed small-scale studies of flow structure, current meters can be lowered from portable bridges. To determine the magnitude of the two horizontal components of velocity, two meters can be positioned at 30–90° to one another. An alternative to deploying two meters is to measure one component of flow with one meter and to use a piece of flexible flagging to determine the net direction of flow and the angle between the meter and the flow.

In addition to the current meter itself, the following equipment is needed for velocity measurement: a support rod or cable, an output device to convert revolutions to velocity and power (usually a battery). A portable computer is often useful for storing information, particularly time series and turbulence information.



(a) (b) (c)

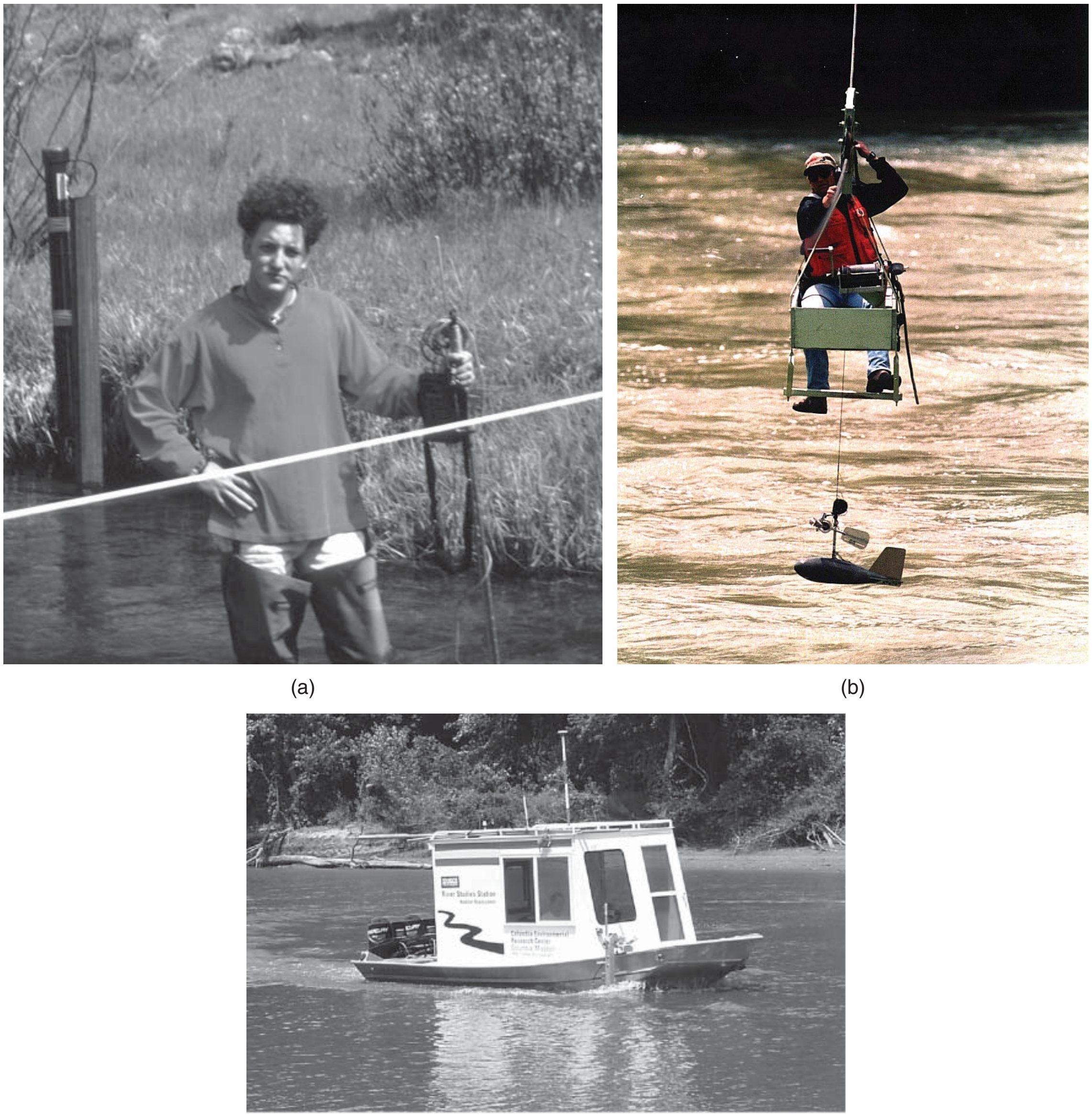
Figure 12.1 Mechanical current meters: (a) standard Price AA current meter – an example of a vertical-axis meter; (b) Price mini-meter – a smaller version of the Price AA meter; (c) Ott current meter – an example of a horizontal axis meter. Source: USGS.

Electromagnetic current meters

Electromagnetic current meters measure flow velocity based upon the Faraday principle that voltage is produced when a conductor (water) moves through a magnetic field produced by the probe. Electrodes on the surface of the probe measure the resulting voltage and the voltage is linearly proportional to the flow velocity.

The use of the electromagnetic current meter has grown in the last decade such that it is now one of more common tools used for measuring flow. The best known of the electromagnetic meters in North America are those manufactured by Marsh-McBirney, such as the Flo-Mate 2000, a 5 cm diameter teardrop-shaped probe with three electrodes (Fig. 12.3). This model provides a measurement of the magnitude of the downstream component of flow over a range from -0.15 to 6 m s^{-1} with an uncertainty of $\pm$2% (Marsh-McBirney 1995). This device attaches to wading rods or cables with the same connector as the Price and mini-current meters. Other models (typically used in oceanographic settings) have 3.7 and 1.3 cm diameter spherical sensors with four electrodes in the horizontal plane, thus allowing the determination of both the downstream and cross-stream components of flow. The current meters have a cosine response to velocity components that are at an angle to the plane of the four electrodes. Both meters can measure bidirectional flow velocities up to $\pm$3 m s^{-1} with an uncertainty of $\pm$2%. In situations where the flow direction may reverse (in separated flow in the lee of bedforms or other obstacles or at depth with stratified flow), the ability of the electromagnetic current meter to measure bidirectional flow can be a distinct advantage over mechanical current meters. This potential limitation in the use of mechanical current meters can be circumvented in clear water with a flexible flag to indicate flow direction. Dinehart (1999) reported that the frequency response of the Marsh-McBirney Flo-Mate 2000 is $\sim$1 Hz or less and therefore does not resolve finer turbulent fluctuations (i.e. $>$2 Hz).

Electromagnetic meters may be affected by strong electrical and magnetic fields and by other electromagnetic current meters placed less than 0.6 m apart, depending upon the model (Marsh-McBirney 1995). The proximity of meters is less of a problem with mechanical current meters.



(a) (b) (c)

Figure 12.2 Modes of deploying current meters: (a) by wading in the stream; (b) suspended from a cableway (photograph courtesy of Post Register Newspaper, Idaho Fall, ID); (c) attached to a boat. Source: Robb Jacobson.

The electromagnetic meter has no mechanical parts, hence it is fairly rugged and is not prone to fouling by debris. Dirt and non-conductive grease and oil should be rinsed from the surface of the probe before storage. Periodically, the zero reading of the meter should be checked. Like the mechanical meters, these meters are robust.

In addition to the current meter itself, the following equipment is needed: a support rod or cable, an output device to convert revolutions to velocity and power (usually a battery). It is common to import the velocity data directly to a data storage device especially if turbulence is being characterized.

Acoustic Doppler velocimeters

Acoustic Doppler velocimeters (ADVs) measure the three components of velocity by the emission of two pairs of acoustic signals from four transducers, their reflection by particles suspended in the flow and the reception of the reflected signal. With the ADV, regions near the sensor are not measurable because there is not enough separation in time between signal emission and reception. Similarly, regions near boundaries are not measurable because the strength of reflections from nearby boundaries swamps the signal of small particles in the flow. A major benefit of such equipment is that there is no

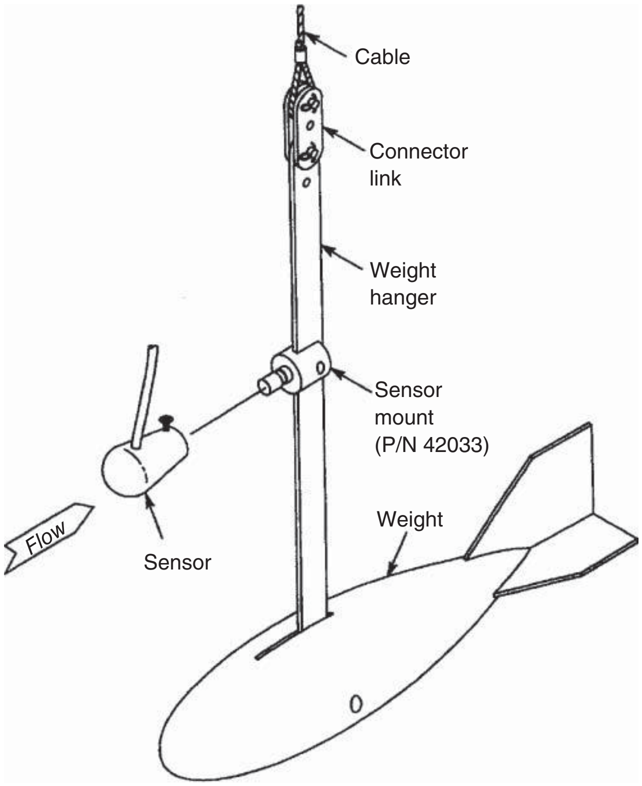


Figure 12.3 Electromagnetic current meter: Flo-Mate 2000 as it would attach to a cable set-up. Diagram courtesy of Marsh McBirney.

Figure 12.4 Two views of an acoustic Doppler velocimeter. The larger view shows the probe and wand and the inset provides a close-up of the wand. Signals are emitted at the base of the wand and are received at the tips of each arm. The total spread of the arms is 77 mm. Photograph courtesy of Sontek, Inc.

device in the sampling volume to distort the flow. Acoustic Doppler velocimetry is useful for characterizing turbulence and turbulent parameters such as the Reynolds stress.

ADVs have become fairly common in the last decade and there exist a variety of types of these devices. Some are small, mounted on a wand with a measuring volume that is relatively small (0.1–0.5 cm^3) and close to the device (5–10 cm) (Fig. 12.4). Such devices are very useful for measuring complex flow fields, especially near boundaries. Others are much larger, with a sampling volume that is large and somewhat distant from the device (programmable from 0.5 to 15 m). The exact specifications vary among manufacturers. The smaller instruments can measure velocities from 0.001 up to 5 m s^{-1} to within several centimetres of bed while some of the larger ones can measure velocities up to 10 m s^{-1}. The uncertainty of measurements is about 0.5–1%. Sampling rates are typically up to 25 or 50 Hz but usually only the 1 s averages are reported. The devices must be moved to obtain information at other locations.

After use, the ADV should be inspected, washed and dried before being stored. Prior to use, all connections should be checked. Otherwise, maintenance of the ADV requires minor effort.

Power (usually batteries) and a data logger or computer to run software and record data are needed to employ an ADV. Some support rod or structure is needed to hold the ADV steady in the flow.

Acoustic Doppler current profilers

Acoustic Doppler current profilers (ADCPs) use the same basic approach as ADVs, but the velocity at multiple points is characterized rather than at a single point. An ADCP measures the three components of velocity at multiple points by the emission of acoustic signals, their reflection by particles suspended in the flow at various distances from the transducer and the reception of the reflected signals from the various particles. Figure 12.5 shows a photograph of such a device. The reflections are separated by time of arrival into uniformly spaced cells for which an average velocity is calculated. The device is deployed at the surface such that signals are sent towards the bed. As with ADVs, ADCPs cannot collect data very close to the device – usually within about 0.2 m of the device – in a layer called the blanking distance. Also, the very strong reflection from the stream bottom, called sidelobe interference, precludes measurement in a region near the bed generally about 0.05–0.1 m thick. As a consequence of these limitations, the device is generally limited to water depths greater than about 0.5 m; even then, only the velocity in cells away from either boundary is determinable. These devices measure velocity in the range 0.00–20 m s^{-1} and have an accuracy of 0.25%. The sampling rate can be as high as 4 Hz.

Meters can be deployed at a single site to provide continuous flow information through the depth profile or towed across the channel to provide a transverse characterization of velocity in addition to the vertical characterization as appropriate for

Figure 12.5 Side and bottom view of an acoustic Doppler current profiler. Photograph courtesy of Robb Jacobson.

discharge determination. The ADCP provides a great deal of information about the interior of the flow. If detailed information is needed near a boundary, an ADV may be more appropriate.

After use, the ADCP should be inspected, washed and dried before being stored. Prior to use, all connections should be checked. Otherwise, maintenance of the ADCP requires minimal effort. The ADCP has a fairly rugged housing; nonetheless, the circuitry means that the equipment should handled carefully. Repair in the field is unlikely.

In addition to the device, the following equipment is needed: a boat (Fig. 12.2c) or support to suspend the ADCP, a computer to run software operating the device and processing data and to record data and power source; a 12 V battery is usually sufficient.

Laser Doppler velocimetry (LDV)

Laser light scattered by small particles during their movement through the sampling volume created by the intersection of beams of laser light can be used to determine the velocity. Light is scattered at frequencies related to the velocity components – the well-known Doppler shift. Two beams of laser light are used to measure a single component of flow and multiple components of the flow can be measured with additional beams. A major benefit of such equipment is that there is no device in the sampling volume to distort the flow. The uncertainty of the measurements is about 0.1%. The sampling volume is smaller than that of an ADV or ADCP: a few tenths of a millimetre on a side. The device must be repositioned to measure velocity at a different location in the flow. There are usually sufficient particles in streamwater to serve as scatterers. The measurement of vertical and downstream components at high frequencies (at least 50 Hz) permits the determination of the Reynolds stress.

This device has seen very little use in rivers and streams, although it has seen some use in oceanographic settings. Given the complexity of the equipment, it is probably most appropriate in the laboratory.

Other velocity measurements

The velocity field at the water surface can be determined by filming (Meselhe *et al.* 1998). Particle image velocimetry (PIV) and large-scale particle image velocimetry (LSPIV) are based on recording the displacement of natural objects in the flow (e.g. foam) or seeded floats that serve as the tracers. MacVicar *et al.* (2012) provided a useful summary of such approaches, including several case studies. Substantial processing and image analysis are required but these techniques are reasonably well developed. Ground-level oblique recording will work, but overhead views from vantage points, tethered balloons, helicopters or aeroplanes are likely to be superior. Such techniques can be useful for quantifying the two-dimensional flow field. The approach can be useful in situations where it is unsafe or impractical to measure flow. In some situations, videography can be cheaper than the cost of hiring a field crew.

If flow is relatively clear, equipment can be submerged in clear housings to film buoyant particles within the flow. These provide two-dimensional velocity information through the water column (Drake *et al.* 1988). In conjunction with another camera, a fully three-dimensional description of flow fields is possible. Another technique uses a thin wire to generate hydrogen bubbles by passing a current through the wire (Schraub *et al.* 1965). These bubbles are then filmed or photographed.

Hot-wire or hot-film anemometry uses the fact that the rate of heat transfer from a solid object is related to the velocity of flow past the object (McQuivey 1973). Although used in the controlled laboratory setting (e.g. Richardson and McQuivey 1968), hot-film anemometry has rarely been used in the field (Grant *et al.* 1968). The calibration is very sensitive to temperature shifts and the probe is prone to breakage and to react with dissolved constituents in the water, thus producing a scale that changes the calibration. With hot-film anemometry, one can measure environmentally common flow velocities of 1–400 cm s^{-1}. Similarly, pitot tubes could be used in the field but are rarely used.

12.3 Discharge measurements

A variety of techniques and equipment exist for measuring flow discharge. In this section, the approaches are described; the principles underlying the approaches are explained; and the accuracy and appropriateness of the approaches are discussed.

An example of the use of discharge measurements comes from my work in gravel-bed streams in Idaho (Whiting *et al.* 1999). I was involved in a water rights case that required knowledge of the streamflow in reaches of channel where the US Forest Service was claiming water. The theory of the case was that sufficient water must flow through the channels to preserve the ability of the channel to move all the bedload over the long term. Knowledge of the streamflow was necessary for several reasons and various descriptors of the streamflow were required. We needed stream discharge because the currency of water rights is the volume of water per unit time. Instantaneous measurements were integrated over time to provide estimates of mean daily flow. We also used instantaneous estimates of flow to associate with concurrent sediment transport measurements in order to build a bedload rating curve. The bedload rating curve multiplied by each daily value of flow over the period of record to give the total flux of sediment was required in the analysis. We also had to determine the annual instantaneous peak flow for flood frequency analysis (see Section 12.5) because the policy of the US Forest Service was that no flow above the 25-year flood would be claimed.

Integration of point measurements

The most common method for measuring flow discharge (Q) is the summation across the channel width (w) of the local products of subsection area (a) and mean flow velocity (u):

$$Q = \sum_{0}^{w} ua \tag{12.1}$$

The mid-section method consists of using the mean velocity as representative of a rectangular area with the dimensions of the measured depth at the point of the velocity measurement and the distance between the two adjacent measurements divided by two (Fig. 12.6). Other methods include averaging adjacent velocity and depth measurements to calculate subsection discharges or accounting for the cross-sectional area enclosed by various velocity contours. Hipolito and Leoureiro (1988) reported that the mid-section method gives the most precise measurements of total discharge through a section. Typically, 20–30 verticals are required across the channel and no subsection should have more than 10% of the flow. A more stringent criterion is that no more than 5% of the discharge is in any vertical. The spacing between verticals is commonly equidistant, but it may be advisable to space verticals to produce segments of equal discharge or to place verticals at breaks in the profile. It is usually necessary to estimate the velocity in the end sections as some fraction of the nearest-bank measurement. Cross-sections where measurements are made should be relatively uniform and in straight reaches without eddies. If possible, sections should not contain vegetation.

The determination of the mean velocity of the subsection is typically made with a current meter. There are several approaches to determining the mean velocity in a vertical.

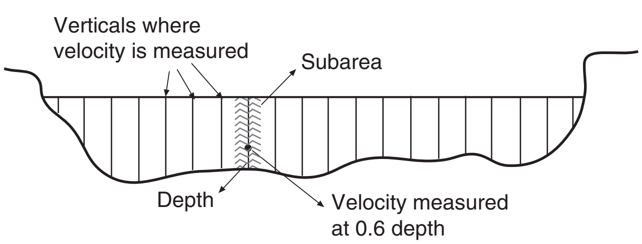


Figure 12.6 Discharge can be determined from the integration the local product of point measurements of velocity and associated area of flow.

Where roughness in the channel is very large relative to water depth and a logarithmic velocity profile may not be typical, velocity can be measured at multiple evenly-spaced points in the vertical (0.1 depth intervals) and the velocities averaged. Where the depths are larger (typically 0.75 m), velocity at 0.2 and 0.8 of the depth below the surface can be averaged. Measurement of velocity at 0.6 of the depth from the surface often gives a good estimate of the mean velocity. This depth corresponds approximately to the elevation of mean velocity given a logarithmic velocity profile. Alternative methods include measurement of the velocity at 0.2 of the flow depth and multiplication of the observation by a coefficient (usually 0.87), and measurement at three points (averaging the mean of measurements at 0.2 and 0.8 of the flow depth with the measurement at 0.6 of the flow depth). Accuracy can be improved by using multiple measurements in the profile, especially near the bed where there often exists the largest gradient of velocity. Velocity should be measured for at least 30 s to account for at least the more frequent pulsation in the flow, especially if the velocity is low.

Carter and Anderson (1963) estimated that the instrumental and sampling error was 4% using a single measurement of velocity at 0.6 of flow depth whereas it was 2.5% using the average of the measurement of velocity at 0.2 and 0.8 of the flow depth; both of these were for a 25-vertical transect. Fulford *et al.* (1994) observed about 2% differences between stream gaugers. The magnitude of the error of the other approaches is unknown.

Earlier in the chapter, the estimation of velocity from floats was described. Ideally, the flow is uniform and the reach straight. Multiplication of the average surface velocity of several floats by a coefficient gives an approximation of the average velocity. The value of the coefficient is typically 0.8–0.9 depending on the resistance to flow. Mosley and McKerchar (1992) suggested a value of 0.86 for the coefficient. Alternatively, the coefficient can be estimated from the Chezy or Manning equations (Herschy 1985). Multiple floats should be used at intervals across the channel to describe the flow field more fully. Although the average velocity determined from several floats can be used to estimate discharge, a better approach is to divide the cross-section into subsections and calculate the discharge through each subsection based on the width, average depth and average velocity in each subsection. These subsection discharges are summed to give the total discharge. The accuracy of such an estimate of discharge is less than with other methods: Herschy (1985) estimated it to be ±10–20%. Such an approach for estimating discharge might be appropriate when other means are too hazardous or when other equipment is not available. If this method is used, the quality of the estimate can be improved with a straight uniform section free of obvious eddies or secondary currents.

Acoustic Doppler current profiling

The acoustic Doppler current profiler (ADCP) uses acoustic signals to measure water velocity and depth. The ADCP transmits short acoustic signals that travel through the water column, strike suspended particles and are reflected back to the receiver.

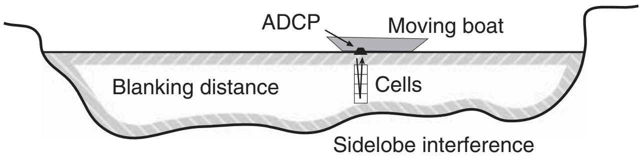


Figure 12.7 Sketch of regions in a channel cross-section analysed by acoustic Doppler current profiling (ADCP). In the hatched areas, velocity measurements are not possible. Near the surface is the blanking distance. Near the bed and banks, this is due to sidelobe interference.

The reflected pulses are separated into stacked 'depth cells' based upon travel time.

For discharge measurement, the ADCP is deployed from a boat (Fig. 12.7). The head of the transducer is submerged just below the water surface and discharge is measured by moving the ADCP across the channel to measure vertical velocity profiles and flow depth. Usually multiple passes across the channel (4–6) are averaged. The vertical velocity profiles generated by the ADCP will not include measurements from near the surface or the bottom. In these regions, acoustic signals have not had a chance to travel a sufficient distance before reflection or the bottom echo makes reflections from small scatters in the flow unrecognizable. Velocities in the unmeasured portions of the profile are estimated using a power-law approximation. Proprietary software to track the bottom, estimate cross-stream position and process the data to yield a discharge value usually comes with the equipment. Additional information on measurement from a boat can be found later in this section.

Morlock (1996) evaluated the device at 12 US Geological Survey gauging stations. The discharge estimated by the profiler was usually close to the conventional estimate; the maximum difference was 8%. The standard deviations of ADCP measurements ranged from about 1 to 6% and were for the most part larger than would be expected from the propagation of errors. Uncertainty in the estimate of velocity is probably about 5%. The device is limited to flow depths greater than 0.5 m. In channels that are 0.5 to 4 m deep, velocities in excess of 2 m s^{-1} are difficult to measure. The ADCP can be very useful in speeding measurements in large rivers.

Rating curves

A common means for estimating discharge is with a simple streamflow rating, typically built from multiple simultaneous measurements of stage and flow discharge (Fig. 12.8). When there is hydraulic control, there will be a unique relationship between the depth of flow and the discharge. The simple rating can be a single curve or a compound curve (set of intersecting single curves) to account for low, moderate and high flows. The rating can be in the form of a table, but more typically is an equation. The general form of most equations is

$$Q = a(G - e)^b \tag{12.2}$$

where Q is the flow discharge, G is the stage, e is typically the height of zero flow and a and b are constants. Common

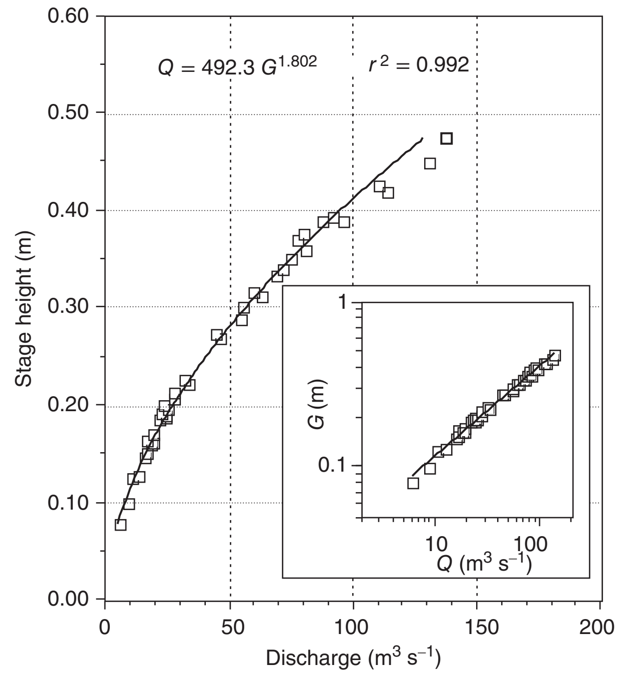


Figure 12.8 An example rating curve from Fourth of July Creek, Idaho. The value of e, the zero offset, is 0.0 m.

values of b range from 1.5 to 2.5 (Herschy 1985). It is usually unnecessary to account for retransformation bias (e.g. Duan 1983). The stage of zero flow can be estimated from the thalweg or by that value which makes the log–log plot linear. The stage is noted visually or monitored by float, bubble gauge (becoming rare), pressure transducer or ultrasonic sensor (described by Herschy 1985) and recorded digitally or by paper trace or punch tape (both now rare). The accuracy of the different sensors varies: ultrasonic (±0.03 m) 0.3% (Latkovich and Leavesley 1992), pressure transducers 0.1–1.0% and float 0.5–1.0%. A stilling well acts to protect the float, pressure transducer or sensor, but more importantly damps vertical fluctuations in the water level associated with wind and flow turbulence. Errors associated with simple ratings are usually less than 5% except at higher flows.

The placement of stream gauges should be done carefully. The best sites are straight reaches of channel without vegetation where flow is confined to a single channel that is not prone to lateral migration or scour and fill. Ideally, a pool exists above the hydraulic control to allow measurement of stage even at low flow. Discharge ratings should be checked at regular intervals – Carter and Davidian (1968) suggested monthly measurement, but more frequent measurement is necessary in streams prone to scour and fill, changing bedforms and vegetative growth. Ratings should be checked after major floods. Discrepancies between the discharge measured and the discharge expected from the stage, if greater than 5%, are usually addressed by subtracting, from the gauge height, a vertical shift that compensates for the discrepancy. Shifts are usually necessary when the bed scours or fills or as other changes in the channel control occur. The datum of all gauges should be checked periodically against several pre-established reference points outside the channel on stable ground.

Complex curves relate discharge to stage and to other variables – such as velocity or the rate of rise or fall of stage – and are used where the stage–discharge relation is complicated by storage or where tidal cycles create variable backwater or reverse flow. Index–velocity ratings can be built by measuring velocity continuously at some specific point in the cross-section using a current meter or across the section using an array(s) of equipment emitting and recording acoustic signals (see the subsection Ultrasonic methods, below). Index–slope ratings can be built by a synchronous array of gauges measuring stage along a reach.

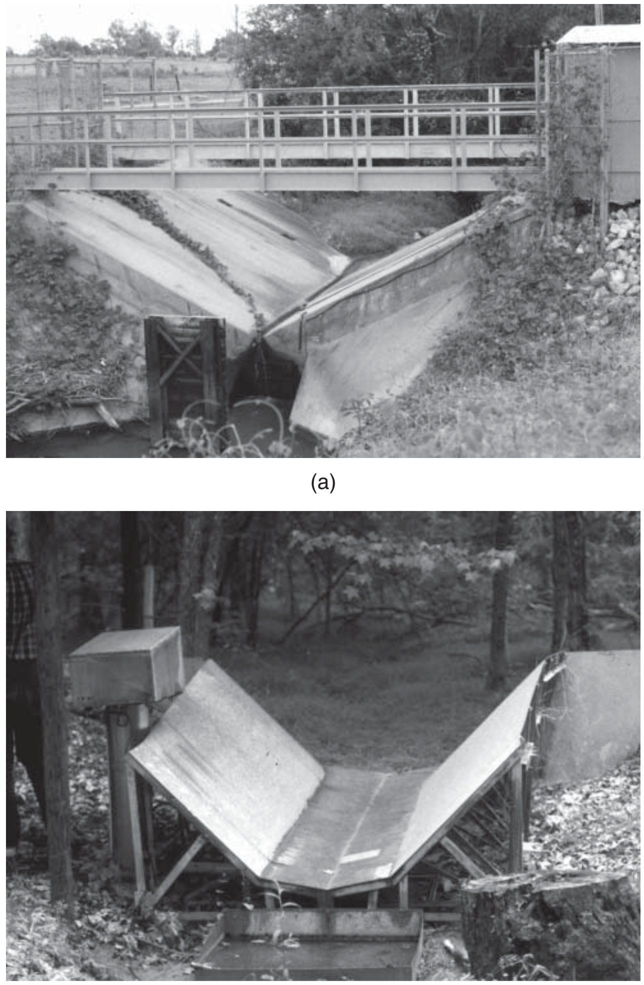
It is often necessary to extrapolate rating curves; this is often less of an issue downward, but can be problematic upward. Herschy (1985) suggested using the Manning equation to estimate discharge by establishing how the quantity $nS^{0.5}$ varied with stage, where n is Mannings' roughness and S is the stream gradient. Alternatively, the rate of increase in average velocity with stage is often very small at high discharge and may approach a constant value. This constant value multiplied by the area of flow provides an estimate of discharge. Nonetheless, extrapolation of the rating curve may provide a reasonable estimate.

Kennedy (1984) provided a detailed description of methods and other issues in discharge rating. Buchanan and Somers (1968) and Herschy (1985) described methods for stage measurement.

Flumes

Flumes have been used to measure discharge in situations where stream characteristics are such that the stage–discharge relationship is prone to shifts due to scour or fill and where the stream is sufficiently small or flashy that other means of gauging are impractical. Most flumes rely on a contraction in the width or a drop in the bed profile to induce critical or supercritical flow in the throat of the flume (Fig. 12.9). Under such conditions, the discharge can be determined by a single depth measurement usually in the throat of the flume because of the unique relation between depth and velocity at critical discharge. Moreover, if the flume is built to specifications, the discharge–head (depth) relation is pre-calibrated and need only be verified by occasional discharge measurements. Other advantages of flumes are that they can operate when head loss is small and over a range of approach velocities.

Flumes can be characterized as subcritical, critical or supercritical. Subcritical flumes are little used at present because flow must be measured in both the approach and the throat of the flume. Critical flow flumes include the Parshall flume and its variants. The Parshall flume is capable of passing small sediment efficiently, thus avoiding clogging the structure and shifting the rating curve. Flumes can be modified with a V-notch. The portable Parshall flume, with a standard throat of 3 in, can be used in settings where discharge is too low for current-meter measurements. Also used for measuring discharge on small watersheds are the HS, H and HL trapezoidal flumes developed by the US Soil Conservation Service. These flumes differ primarily in their capacity. The supercritical flumes are installed typically where sediment is likely to accumulate in the structure unless very high velocities can be used to preclude sedimentation. All flumes can be equipped with a stage recorder to provide a continuous record of stream discharge. Supercritical flumes can have vertical sides (i.e. San Dimas flumes) or trapezoidal sides. Herschy (1985) suggested that the error in estimates of discharge using a flume is 2–4%.



(a)

(b)

Figure 12.9 Flumes may be used in various settings and constructed of various materials. (a) A concrete flume associated with a bridge and a pump house; (b) a flume constructed of sheet metal and wood. Both flumes are in the Goodwin Creek watershed in Mississippi.

Weirs

Weirs are amongst the oldest, simplest and most reliable means for measuring discharge. A weir temporarily ponds streamflow prior to its spilling in freefall over a controlled outlet (Fig. 12.10). If a standard geometry is used, the weir is pre-calibrated for

Figure 12.10 V-notch weir. Photograph courtesy of US Geological Survey.

discharge. Weir blades are either sharp- or broad-crested. Weirs can be rectangular, trapezoidal or have a 90° V-notch. The height of water over the weir blade (the height and the length of the weir crest in the case of rectangular weirs) is measured to estimate the discharge. There is a trade-off between accuracy and maintenance between the two types of crests. The sharp-crested weir is more prone to clogging by floating debris, but is more accurate. The elevation of the water in the pond can be measured continuously and, in conjunction with a rating curve, provides a continuous record of discharge. Whereas debris screens can be installed upstream of weirs to minimize accumulation of floating debris or sediment in the weir, a flume, in particular a supercritical flume, should be used if substantial amounts of sediment or debris are expected. Herschy (1985) suggested that the error in estimates of discharge using a weir is up to 3%. Kilpatrick and Schneider (1983) provided a fairly complete summary of the use of flumes and weirs for measuring discharge.

Ultrasonic methods

This less common technique is based upon the continuous measurement of the time of travel of sound waves emitted from transmitters deployed below the water surface. The time of travel over the distance between transmitter and receiver is related to the average flow velocity, which multiplied by the flow area associated with the velocity gives the discharge. The elevation in the water column at which the velocity is measured is fixed, so the relation between velocity at the measurement height and stage must be established. An array of transducers can be deployed to measure the velocity at various heights in the water column (e.g. Genthe and Yamamoto 1974). Total discharge is calculated as the sum of the product of average velocity at a given height and the cross-sectional area associated with the velocity measurement. Such an arrangement may be useful where the range in stage is large or greater precision in discharge estimates is needed. The velocity may be determined to 0.1% (Herschy 1985) but the uncertainty of a discharge estimate based upon such a measurement is approximately 5% (Herschy 1985).

The method may be useful where no stable stage–discharge relationship exists and the construction of a weir or flume is not feasible. Such an approach can be used in backwater from tides, downstream tributaries or dams.

Dilution and tracer gauging

The dilution method for determining flow discharge involves the addition of a conservative tracer to the flow and the determination of the concentration of the tracer downstream after the tracer has been well mixed throughout the flow. In the constant-rate injection method, the tracer of concentration, C_1, is applied at a constant measured rate q. Downstream, where the concentration is C_2, the discharge (Q) is

$$Q = \frac{C_1}{C_2} q \tag{12.3}$$

When using continuous tracer injection, the downstream measurement point must be sufficiently far downstream that the tracer is well mixed across the channel and at depth. An alternative approach is to introduce a slug of tracer and to monitor the concentration at a downstream location. The discharge is

$$Q = \frac{V}{T} \frac{C_1}{\overline{C_2}} \tag{12.4}$$

where V is the initial volume of the tracer, T is the time of passage for the tracer (first arrival to last arrival) and $\overline{C_2}$ is the average concentration over the time of passage. Hubbard *et al.* (1982) measured the discharge by tracking a slug of tracer. The average velocity of the centroid of the dye cloud should approximate the mean flow velocity, which when multiplied by the cross-sectional area gives the discharge.

The tracer for use in dilution gauging can be a dye or solute. Radioactive tracers were used in the past but are generally not used nowadays. Fluorescent dyes have been used as tracers to determine discharge (Wilson 1968). Fluorescence varies with concentration, which in turn varies with discharge. Factors affecting the relationship between fluorescence and concentration and in turn discharge include temperature, pH, reactions with other constituents and photochemical degradation of the dye. Concentrations can be detected with a fluorimeter to below 100 ng L^{-1}. When solutes are used, sodium chloride is the most common. It is not generally a problem if the tracer already exists in the stream; if the background concentration remains steady, the dilution method can still be used. If salts are used in the tracing, the concentration of dissolved salts is linearly related to electrical conductivity, which is relatively easy to monitor.

Tracer dilution may be useful in various situations, but is particularly useful where flow is shallow or clogged with vegetation or other debris or where the lateral input of water is large (many tributaries and/or seepage) and the change in discharge must be known through the reach. Herschy (1985) reported that dilution gauging has been used with discharges as large as 2000 $m^3 s^{-1}$. As noted in the Introduction, the technique is predicated upon the flow being well mixed at the sampling location. The uncertainty is about 5%.

More detailed discussions of the use of dye or salts in dilution and tracer gauging are available in Wilson (1968), Hubbard *et al.* (1982) and US Bureau of Reclamation (1984).

'Moving-boat' method

The measurement of discharge by the moving-boat method may be practical for wide rivers or remote locations, or when conditions are unsteady (e.g. tidally influenced) or hazardous such that measurements need to be made rapidly. Discharge is measured by equipping a boat with a depth sounder and a continuously operated current meter. While traversing the stream, the depth and combined stream and boat velocities and angle of the flow with respect to the traverse are measured. The current meter is set at some characteristic depth below the surface and the measured velocity correlated with the average velocity. The value of the coefficient is typically 0.87–0.96 for velocities measured 1 m below the surface (Herschy 1985). The minimum speed of the boat should be such that it traverses the river in a straight line roughly orthogonal to the flow; this will require that the boat point at an angle of 20–60° (upstream) to the cross-stream direction (Herschy 1985). The mid-section method for integrating subsection measurements to yield total discharge is recommended. For the best measurements, traverses should contain 30–40 observation points and the discharge from six or more traverses should be averaged. Smoot and Novak (1969) estimated that discharges by this technique are within 5% of measurements by conventional techniques.

The ACDP approach described earlier relies upon the moving-boat method. The book by Herschy (1985) includes a chapter devoted to moving-boat measurements.

Electromagnetic method

Electromagnetic determination of streamflow discharge has been accomplished in some settings. The flow of water in a stream through an electromagnetic field induces an electromotive force that is measured and is directly related to the average flow velocity in the cross-section. The earth's electromagnetic field is useful in principle for such measurements, but electrical interference is a problem. Typically, a coil is buried in the streambed through which electric current is passed and the resulting electromagnetic field is used to measure discharge. The equipment is relatively expensive to install, but in small streams where the flow can be passed through a pipe the costs can be more reasonable. Usually AC current is needed. Electromagnetic discharge measurement can be used to determine discharge when average velocities are as low as 0.2 cm s^{-1} (Herschy, 1985).

Correlation of point measurements with discharge

Point measurements of flow velocity or average velocity in a vertical have been used to estimate discharge. For instance, acoustic Doppler velocimeters can be placed on the stream bottom looking upwards, moored looking downwards or mounted on some structure looking sideways to provide a measure of velocity in some defined region at a distance of 0.5–2 m from the device. Alternatively, acoustic Doppler current profilers can be attached to the bottom or some other structure or moored and the average velocity determined for the vertical (Williams 1996). These local measurements of velocity are correlated with measured stream flow, much like a rating curve, to estimate flow discharge. The accuracy of such correlation methods is probably no better than 5%.

Another correlation approach is to use a hydraulic model and stage data and occasional measurements of surface velocity to estimate flow discharge. Corato *et al.* (2011) described such an approach and suggested errors of no more than 5%.

Other techniques for discharge determination

A new class of techniques for measuring flow without contacting the water surface has been developed in the last decade or so. These techniques typically use radar (Costa *et al.* 2006) or particle image velocimetry to characterize the surface velocity field (Creutin *et al.* 2003; Muste *et al.* 2008). Such techniques offer the important benefit of being useful in high flows and for making measurements in remote areas. An issue to consider when using surface velocity measurements to quantify discharge is the appropriate value of the conversion factor between surface velocity and mean velocity in the water column. Costa *et al.* (2000) suggested 0.85 and others have adopted this value (e.g. Muste *et al.* 2008), but some have suggested that other values are more appropriate. MacVicar *et al.* (2012) found that the use of a value of 0.90 better matched discharges measured by other methods.

In some settings, discharge can be determined by measuring the volume or mass of water collected over a specific time interval. Volumetric measurement of the freefall of water is a convenient way to verify flume and weir calibration. These methods are most useful for small discharges (up to a few litres per second).

12.4 Indirect methods of discharge estimation

Several indirect approaches exist for estimating flows when other techniques are not suitable or available. For instance, it is often necessary to estimate the magnitude of streamflow where the best evidence is high water marks – mud lines, deposited sediment and/or debris lines. Other records may not exist because there was no gauge in the reach of channel, existing equipment was inoperable or destroyed by the flows or access by personnel to the sites during the high flow was impossible or unsafe.

Slope–area method

One such indirect method for discharge estimation is the slope–area method (sometimes called the normal flow equations). It is based upon resistance equations and the cross-sectional area and slope of the channel and it assumes uniform steady flow. The best known of these methods, at least in the United States, is the Manning equation:

$$Q = \frac{AR^{\frac{2}{3}}S^{\frac{1}{2}}}{n} \tag{12.5}$$

which is similar in form to the Chezy equation:

$$Q = cA(RS)^{\frac{1}{2}} \tag{12.6}$$

The slope (S) and the hydraulic radius (R) of the channel should be measured in as uniform a reach as is possible. The area of flow is A and n and c are roughness and conveyance factors, respectively. The slope should be measured over a length of 20 or more channel widths unless such a distance would include major discontinuities in width or depth or falls. The value of the roughness factor, n, can be determined by calibration, from empirical relations or by comparison with descriptive tables or to a reference atlas (Barnes 1967). The roughness factor depends on the size of the bed material, the presence of bedforms and the amount of vegetation. For most channels, it varies from about 0.01 to 0.06, but values of 0.1 and greater are observed in channels with boulders and other large roughness elements (Hicks and Mason 1991). In compound channels or with flow over a floodplain, different roughness values can be ascribed to various subsections of the cross-section. Discharge estimated by the slope–area method has an uncertainty of at least 10–20% (Herschy 1985).

Dalrymple and Benson (1967) explained the use of high water marks for estimating peak flows by the slope area method.

Contraction method

Another indirect method includes the use of the energy equation. At channel width contractions, such as those created by bridge abutments, the change in the water surface elevation through the contraction can be used to estimate peak discharge. The water surface elevation could be measured during high flow but is more often taken from high water marks. One issue that needs to be considered is the possibility of scour and fill of the bed during high flow (Matthai 1967). Peak discharges can be determined also from high water marks of the headwater and tailwaters above and below culverts. The approach can be used for subcritical and critical flows, transitions between such flows and submerged outlets (Bodhaine 1968). Uncertainties associated with estimates of peak flow by these methods are probably 20% or greater.

Step-backwater modelling

A final suggestion for estimating discharge in the absence of measurement is the use of step-backwater models for computing the discharge associated with observed water surface profiles as summarized by Miller and Cluer (1998). Various hydraulic models based on the one-dimensional energy equation are available that use an iterative solution technique known as the step-backwater method. The hydraulic package HEC-2 (Hydrologic Engineering Center 1982) is an example of such a model. To estimate flows, bathymetry and estimates of roughness are necessary. The discharge that best matches the observed water surface elevations (often high water marks) along the channel is the estimated streamflow. Uncertainties are relatively large – probably at least 10–20%.

Such an approach can be particularly useful for estimating the magnitude of exceptionally large floods on ungagged streams or on gauged streams when the flood has destroyed the measurement equipment. Step-backwater modelling can also be used to establish the magnitude of palaeofloods when suitable high water markers can be found (e.g. Beebee and O'Connor 2003).

12.5 Flow hydrographs and analysis of flow records

The measurement of velocity and discharge is important in its own right, but these measurements are often used to develop a hydrograph describing the flow rate over time (Fig. 12.11). Hydrographs can be used to correlate runoff timing and volume to precipitation timing, intensity and duration or to determine 1-, 7-, 28-day, etc, high or low flows among a few examples. Instead of the history of flows, the distribution of flows can be analysed (Fig. 12.12).

At the beginning of the section on discharge measurement, the example of a water rights case was presented. The history of flows was used to answer various questions about the effect of the US Forest Services claims for water – How much water would be claimed? During which months? How many years would there be no Forest Service claim because flow levels did not rise high enough?

In the next few subsections, the presentation and analysis of flow records are discussed.

Flow hydrograph

In Section 12.3, a variety of methods were suggested for determining discharge. Although in principle almost any of these methods could be repeated with great frequency to provide hydrograph of streamflows, it is most common to use a stage–discharge rating curve and the history of stage to produce the hydrograph (Fig. 12.11). Stage is commonly measured at 15 minute intervals and recorded, the stage converted to discharge and then the discharge recorded.

In the United States, stream flow information is available from the United States Geological Survey's National Water Information System (waterdata.usgs.gov/nwis). At about 26,000 sites, surface water quantity measurements are made. The collected daily information includes mean, median, maximum and minimum values. Statistics of daily mean time series for daily, monthly and annual time periods are also available, as are annual maximum instantaneous peak steam flows and gauge heights. At a subset of these sites, real-time data (15–60 minute

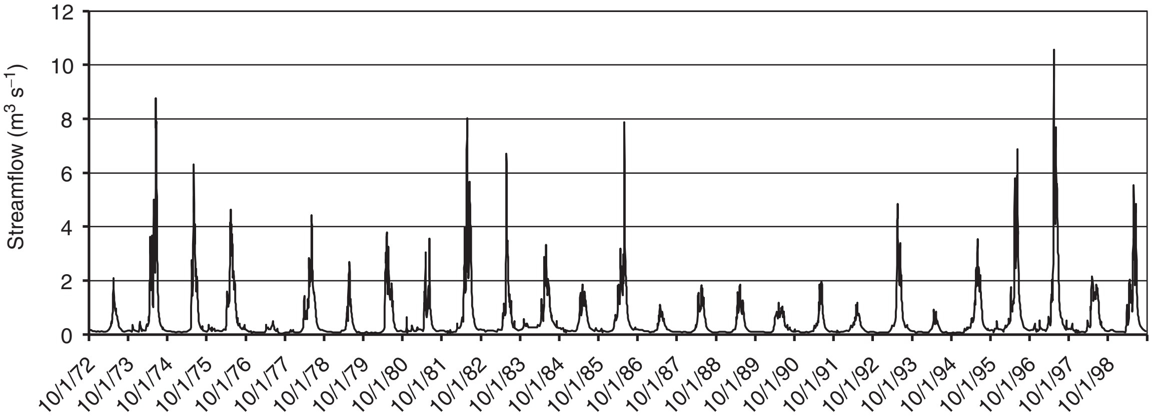


Figure 12.11 Flow hydrograph for Thompson Creek, Idaho, for the period 1 October 1972 to 30 September 1999.

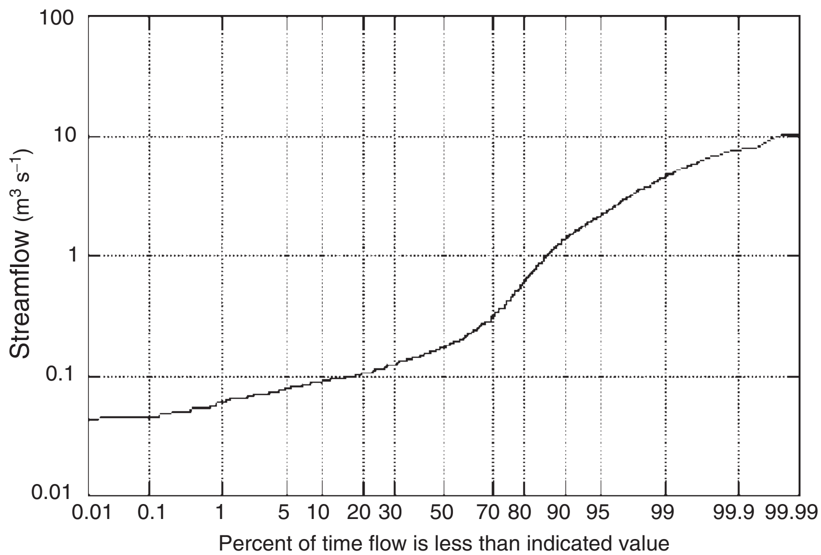


Figure 12.12 Flow duration curve for Thompson Creek, Idaho, for the period 1 October 1972 to 30 September 1999. The discretization period is daily.

intervals) are transmitted to the database at 1–4 hour intervals and are available. The Water Survey of Canada has similar data, as does the National River Flow Archive in the United Kingdom and the hydrobank in France (http://hydro.eaufrance.fr/). It is also possible to have online information on discharge at hour intervals, as is shown at http://www.rdbrmc.com/hydroreel2/station.php?codestation=6 for the Ain River, France.

Flow duration curves

Flow duration curves (Fig. 12.12) describe the percentage of the time flow is greater than a specified value (percent exceedence). An important consideration in such cases is the time interval of measurement (discretization). In basins with short lag times (small, urban and/or extensive bare rock), the mean daily flow is a poor descriptor of the observed flow. For such basins, the appropriate interval with which to build flow duration curves may be 15 minutes or shorter. In many cases, shorter intervals were used to determine the mean daily flow, but it may be necessary to return to primary records (digital files, punched tapes, hydrograph traces) to recover the finer time resolution. The flow duration curve is developed by sorting the average discharge over the selected discretization interval and assigning the appropriate probability of exceedence based on the total length of the record. It is critical that there be no missing values over the period of record. If data are missing, they must be estimated. Frequency analyses should be avoided with records shorter than 10 years or for estimating the frequency of events greater than twice the record length (Viessman *et al.* 1977).

Extreme value plots

For many purposes, extremes of the streamflow (high and low) are critical information. Peak instantaneous discharge is the typically determined for a gauge to characterize the recurrence interval of floods (Fig. 12.13). Usually the annual peak over

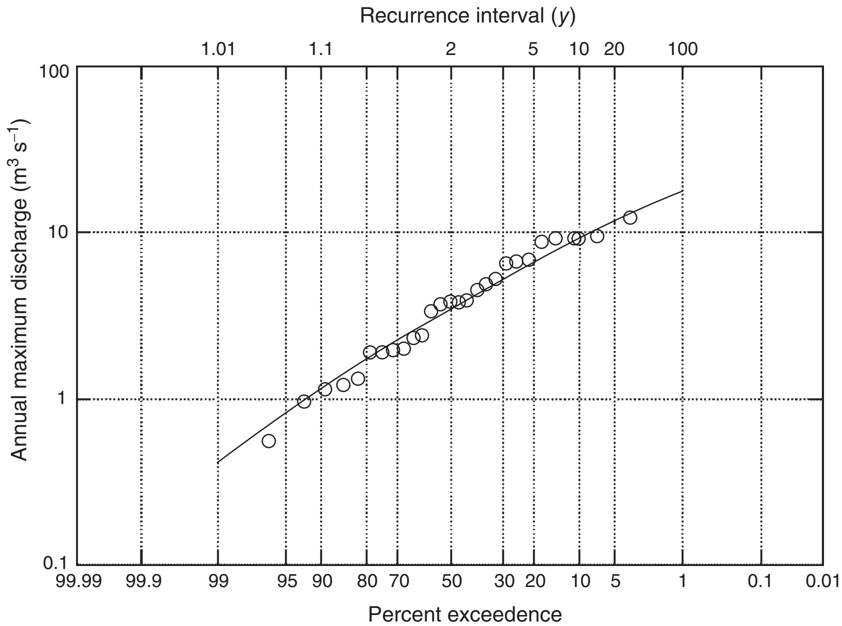


Figure 12.13 The frequency and recurrence interval of annual peak floods for Thompson Creek, Idaho, for the period 1 October 1972 to 30 September 1999.

the period of record is determined, ranked by magnitude and, depending on the record length, the probability of exceedence is determined. The inverse of the probability of exceedence is the recurrence interval. Instead of analysing the peak flow of each year, which is called an annual series, a partial series including all peaks above some threshold can be analysed. Annual low flows (e.g. 1-, 7-, 28-day) are very important for water supply and environmental studies and can be analysed similarly.

These tools are treated extensively in hydrology textbooks such as those by Dunne and Leopold (1978) and Dingman (2002).

12.6 Issues in selecting methods

As shown above, there are many approaches and technologies that may be appropriate for characterizing flow and flow velocity and determining discharge. The issues to consider in the selection of an approach or equipment are numerous but might be categorized as follows:

- the purpose of the measurements;
- appropriateness of pre-existing data;
- the required precision and accuracy;
- the channel attributes (size and geometry of the channel, stability of the reach);
- the hydrological attributes (steadiness of flow, unidirectional or reversing flow);
- site accessibility and infrastructure for making measurements;
- the equipment available;
- time available to make measurements;
- cost (equipment and personnel time).

The following sections elaborate briefly upon these issues.

Purpose of measurements

There are a variety of reasons for collecting flow information, a number of which were posed in the Introduction and include monitoring (e.g. the amount of water), basic research (e.g. the turbulence associated with sediment motion) and applied research (e.g. the amount of habitat available at different flow levels). Depending upon the purpose for which flow information is collected, various methods may or may not be suitable. An important question to consider is the level of spatial and temporal detail required. If spatial detail is required, near steady flow for long periods is helpful. Snowmelt-driven systems, spring-fed streams and streams with reservoir releases are likely to maintain relatively high flows near formative conditions, which is especially important for detailed sediment transport studies. Measurements near base flow may provide opportunities for detailed measurement but at stages far different than the flows that have the most influence on channel form.

Pre-existing data

A number of governmental and non-governmental agencies collect flow information that may be suitable for the problem at hand. In other situations, pre-existing data may be a useful starting point. For example, a discontinued gauging station could be re-occupied. If the rating curve could be shown to be still valid with a few measurements of stage and discharge, a great deal of effort could be avoided. If nothing else, pre-existing data on the stream or nearby streams may suggest the sort of flows expected or the timing of flows, thus aiding in the experimental design.

It is often the case that only the processed mean daily discharge values and instantaneous peaks are published. Sometimes it is possible to retrieve more detailed flow information [cross-section data, hydrograph chart traces (digital or paper)] by contacting the agency that collected the information.

Precision and accuracy

Depending in part on the purpose of the measurement, various levels of precision and accuracy may be required. For instance, if the question being addressed is the change in runoff volume associated with a small change in impervious area and the expected change is about 10%, it makes sense to use a methodology with an uncertainty substantially less than 10%. While many sets of discharge measurement using a technique with larger uncertainty have the potential to be suitable, the effort required is greater. In other situations, the flow or discharge may need to be known very accurately because of the importance or costs of decisions based upon such information.

The uncertainties in velocity measurement as summarized from the earlier discussion are given in Table 12.1. The uncertainties in discharge measurement (following Herschy 1985) for moderate flow conditions are given in Table 12.2. Uncertainties are likely to be larger at very high and very low flows. Uncertainties in mean daily, mean monthly and mean annual discharge will be lower.

The uncertainties outlined in Tables 12.1 and 12.2 should be taken as approximate estimates. Factors that can affect the magnitude of the uncertainty are the training and care of the operator, the condition of the equipment, the precision and accuracy of the instrumentation, the number of verticals in the section, the precision of depth measurement, the measurement of stage, the stage–discharge relation, the steadiness and uniformity of flow in the measurement reach, the geometry in the measurement reach, the relative amount of unmeasured flow and other factors.

Although not universally true, there is some truth in the generalization that more accurate measurements require more expensive equipment, more time and more personnel.

Table 12.1 Uncertainty in velocity measurement.

Method	Uncertainty (± %)
Floats	10–20
Mechanical current meters	0.5–2
Electromagnetic current meters	~2
Acoustic Doppler velocimeters	0.5-1
Laser Doppler velocimeters	~0.1

Table 12.2 Uncertainty in discharge measurement.

Method	Uncertainty (± %)
Floats	10–20
Integration of point measurements	~5
ADCP	5
Rating curve	<5
Flume	2–4
Weir	1–3
Ultrasonic	~5
Dilution	5
Moving boat	~5

Channel attributes

Channel attributes include the size and geometry of the channel, the nature and size of the substrate and the stability of the reach. The size of a system can influence the type of equipment and methods appropriate for use. Discharge from small streams may be measured best with weirs or flumes whereas the largest rivers may be measured best with a moving boat – equipped with current meters or an acoustic Doppler current profiler. The nature and size of the substrate can be important. The size of the sediment on the stream bottom can influence the measurement approach in several ways. If the size of the particles on the bed is large relative to the flow depth, the velocity profile may be non-logarithmic, hence measuring velocity at a single elevation at 6/10 the depth will not be appropriate. If the relative roughness is large, techniques relying upon the transmission of a signal and its reflectance may give spurious results. On the other hand, fine beds will often develop bedforms potentially requiring longer averaging periods to account for migration of bed features. Where flow accelerations are large, as for example at Solfatara Creek mentioned earlier, the measurement of discharge will require multiple measurement points in the vertical. The presence of in-channel vegetation can affect measurement. Vegetation can clog mechanical current meters whereas the vegetation does not affect electromagnetic current meters.

The stability of the reach can be critical for certain techniques of discharge determination. Discharge, as determined by the building of a rating curve, requires that there is a consistent relationship between stage and volume of flow. If the bed is aggrading or degrading, this requirement is not met. One of the other techniques would have to be used in such situations.

Hydrological attributes

Hydrological attributes to consider in the selection of methods include whether flow varies periodically (as with diurnal snowmelt), whether the response is rapid (as with a small urbanized basin) and whether flow is unidirectional or reverses (as with tides). Systems prone to rapid changes in flow generally require equipment that collects information automatically. For example, where there is a diurnal signal, it may be warranted to measure flow at a consistent time; where runoff occurs rapidly, stage could be recorded by a pressure transducer and converted to discharge; and in reaches affected by tides, flow reversals may require that meters are capable of indicating the direction of flow.

The range of flows to be measured may also affect the selection of equipment and methods. For instance, there are situations where discharge measurement by wading at low flow is possible (even preferable), but at high flow suspension of equipment from a bridge or cables is required. Should a single approach be used at all flows or should different measurement approaches be used at different stages? There is not a simple answer to the question: the relative merits of collecting the best information over a particular range of flows and using a single approach must be balanced.

Site accessibility and infrastructure for making measurements

The accessibility and remoteness of sites may influence the selection of methods and equipment. If a site is remote, it may be impractical to rely primarily upon personnel to collect flow measurements or to reach the site in a timely manner. If a site is not accessible by a road or easily accessible by boat, measurements may need to rely upon equipment that collects data automatically. If equipment must be carried overland, the weight of equipment and peripherals becomes a consideration.

The infrastructure at a site or site conditions can affect the approach selected. The presence of electric power makes it easier to use electromagnetic discharge determination techniques and ultrasonic techniques. A bridge spanning the river may make a site suitable for suspending current meters (even though they may affect flow) whereas the absence of the bridge may require working from a boat.

Some sites may be inaccessible at certain times of the year. Breeding grounds or spawning habitat of endangered or threatened species often prevent personnel from visiting sites to collect data or download automatically collected data. In these situations, automatic data collection and either large data storage ability or data transmission capability are necessary.

Equipment

It is not unusual for the availability of equipment or familiarity with a particular type of equipment to affect the design of the data collection plan. The robustness of equipment and ease of operation are important to consider, as is the weight and the amount of peripheral equipment, especially if the site is remote and walking into the site or boat access is required.

Time

The time required to make flow and discharge measurements varies appreciably by method and by equipment. In addition, the time for data processing can also be appreciable.

In some situations the speed at which measurements can be made is important. If debris is in the channel or if there is boat traffic, time can be of the essence. In situations where flow velocity or discharge is changing rapidly, the ability to make a measurement in a short time interval is important. As an example, characterizing the flow field over a bedform from the bed to the water surface should be accomplished very rapidly – before the bedform migrates any appreciable distance. The time required to make a measurement and process data will probably be considered primarily in the context of the purpose of the measurement, the required accuracy and the cost.

Cost

The expenditure of funds in the purchase of equipment for flow measurement can vary by an order of magnitude at least. Some variation in cost (and quality) of largely similar equipment exists between manufacturers. Additional costs are associated with the collection and with the processing of data. A modest sum should be reserved for the maintenance of equipment.

12.7 Conclusion

As laid out in earlier parts of this chapter, there are a variety of reasons for making measurements of flow, a variety of equipment and methods available and a variety of reasons for selecting one particular approach. This primer on flow measurement should not be taken as sufficiently detailed to serve as a stand-alone guideline on any particular method or equipment. Perhaps the most appropriate use of the primer is in the initial phases of investigative design. It is recommended that the hydrographer who is considering one of the methods outlined herein read more detailed descriptions as suggested at the end of Section 12.1 or listed in the references.

References

Barnes, H.H. 1967. *Roughness Characteristics of Natural Channels, US Geological Survey Water Supply Paper 1849*, Washington, DC: US Government Printing Office, 213pp.

Beebee, R.A. and O'Connor, J.E. 2003. The Outhouse flood: a large Holocene flood on the Lower Deschutes River, Oregon. In: O'Connor, J.E. and Grant, G.E., eds., *A Peculiar River: Geology, Geomorphology and Hydrology of the Deschutes River, Oregon*, Washington, DC: American Geophysical Union.

Bodhaine, G.L. 1968. Measurement of peak discharge at culverts by indirect methods. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A3.

Buchanan, T.J. and Somers, W.P. 1968. Stage measurement at gaging stations. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A7.

Buchanan, T.J. and Somers, W.P. 1969. Discharge measurements at gaging stations. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A8.

Bureau of Reclamation. 1984. *Water Measurement Manual*. Washington, DC: US Government Printing Office, 327pp.

Carter, R.W. and Anderson, I.E. 1963. Accuracy of current-meter measurements. *American Society of Civil Engineers Journal* 89: 105–115.

Carter, R.W. and Davidian. J. 1968. General procedure for gaging streams. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A6.

Christensen, B.A., 1994. Velocity measurements by the "one-orange method". In: Pugh, C.A., ed., *Fundamentals and Advancements* *in Hydraulic Measurements and Experimentation. Proceedings of the Symposium on Fundamentals and Advancements in Hydraulic Measurements and Experimentation, 1994, Buffalo, NY*, New York: American Society of Civil Engineers, pp. 76–85.

Corato, G, Moramarco, T. and Tucciarelli, T. 2011. Discharge estimation combining flow routing and occasional measurements of velocity. *Hydrology and Earth System Sciences* 15: 2979–2994.

Costa, J.E., Spicer, K.R., Cheng, R.T., Haeni, F.P., Melcher, N.B., Thurman, E.M., Plant, W.J. and Keller W.C. 2000. Measuring stream discharge by non-contact methods: a proof-of-concept experiment. *Geophysical Research Letters* 27: 553–556.

Costa, J.E., Cheng, R.T., Haeni, F.P., Melcher, N., Spicer, K.R., Hayes, E., Plant, W., Hayes, K., Teague, C. and Barrick, D. 2006. Use of radars to monitor stream discharge by noncontact methods. *Water Resources Research* 42: W07422.

Creutin, J.D., Muste, M., Bradley, A.A., Kim, S.C. and Kruger, A. 2003. River gauging using PIV techniques: a proof of concept experiment on the Iowa River. *Journal of Hydrology* 277: 182–194.

Dalrymple, T. and Benson, M.A. 1967. Measurement of peak discharge by the slope–area method. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A2.

Dingman, S.L. 2002. *Physical Hydrology*, 2nd edn., New York: Macmillan, 600pp.

Dinehart, R.L. 1999. Correlative velocity fluctuations over a gravel bed river. *Water Resources Research* 35: 569–582.

Drake, T.G., Shreve, R.L., Dietrich, W.E., Whiting, P.J. and Leopold, L.B. 1988. Bedload transport of fine gravel observed by motion-picture photography. *Journal of Fluid Mechanics* 192: 193–217.

Duan, N. 1983. Smearing estimate: a nonparametric retransformation method. *Journal of the American Statistical Association* 78: 605–610.

Dunne, T. and Leopold, L.B. 1978. *Water in Environmental Planning*, San Francisco, CA: W.H. Freeman, 818pp.

Fulford, J.M., Thibodeaux, K.G. and Kaehrle, W.R. 1994. Comparison of current meters used for stream gaging. In: Pugh C.A., ed., *Fundamentals and Advancements in Hydraulic Measurements and Experimentation. Proceedings of the Symposium on Fundamentals and Advancements in Hydraulic Measurements and Experimentation, 1994, Buffalo, NY*, New York: American Society of Civil Engineers, pp. 376–385.

Genthe, W.K. and Yamamoto, M. 1974. A new ultrasonic flowmeter for flows in large conduits and open channels. In: Wendt, W.E., ed., *Flow: Its Measurement and Control in Science and Industry*, Pittsburgh, PA: Instrument Society of America, pp. 947–955.

Grant, H.L., Stewart, R.W. and Moilliet, A. 1968. The spectrum of temperature fluctuations in turbulent flow. *Journal of Fluid Mechanics* 34: 423–442.

Herschy, R.W. 1985. *Streamflow Measurement*, London: Elsevier Applied Science, 553pp.

Hicks, D.M. and Mason, P.D. 1991. *Roughness Characteristics of New Zealand Rivers*, Christchurch: Water Resources Survey, DSIR Marine and Freshwater.

Hipolito, J.N. and Leoureiro, J.M. 1988. Analysis of some velocity-area methods for calculating open-channel flow. *Hydrological Sciences Journal* 33: 311–320.

Hubbard, E.F., Kilpatrick, F.A., Martens, L.A. and Wilson, J.F. 1982. Measurement of time of travel and dispersion in streams by dye tracing.In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A9.

Hydrologic Engineering Center. 1982. *HEC-2 Water Surface Profiles – Users Manual*, Davis, CA: US Army Corps of Engineers.

ISO. 1983. Measurement of Liquid Flow in Open Channels, *ISO Standards Handbook*, Geneva, International Organization for Standardization, 518pp.

Jarrett, R.D. 1992. Hydraulics of mountain rivers. In: Yen, B.C., ed., *Channel Flow Resistance: Centennial of Manning's Formula*, Littleton, CO: Water Resources Publications, pp. 287–298.

Kennedy, E.J. 1984. Discharge ratings at gaging stations. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A10.

Kilpatrick, F.A. and Schneider, V.R. 1983. Use of flumes in measuring discharge. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A14.

Latkovich, V.J. and Leavesley, G.H. 1992. Automated data acquisition and transmission. In: Maidmont, D.R., ed., *Handbook of Hydrology*, New York: McGraw-Hill, pp. 25.1–25.21.

MacVicar, B.J., Hauet, A., Bergeron, N., Tougne, L. and Ali, I. 2012. River monitoring with ground-based videography. In: Carboneau, P.E. and Piégay, H., eds., *Fluvial Remote Sensing for Science and Management*, Chichester: John Wiley & Sons, pp. 367–384.

Marsh-McBirney. 1995. *Operation and Maintenance – Electromagnetic Current Meters*, Frederick, MD: Marsh-McBirney, Inc., 116pp.

Matthai, H.F. 1967. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A4.

McQuivey, R.S. 1973. *Principles and Measuring Techniques of Turbulence Characteristics in Open-channel Flows, Geological Survey Professional Paper 802-A*, Washington, DC: US Government Printing Office, 82pp.

Meselhe, E.A., Bradley, A.A., Kruger, A. and Muste, M.V.I. 1998. PIV and numerical modeling for flow analysis. In: *Water Resources Engineering '98. ASCE International Water Resources Engineering Conference Proceedings*, vol. 1, pp. 526–531.

Miller, A.J. and Cluer, B.L. 1998. Modeling considerations for simulation of flow in bedrock channels. In: Tinkler, K.J. and Wohl, E.E., eds., *Rivers Over Rock: Fluvial Processes in Bedrock Channels, AGU Geophysical Monograph 107*, Washington, DC: American Geophysical Union, pp. 61–104.

Morlock, S.E. 1996. *Evaluation of Acoustic Doppler Current Profiler Measurements of River Discharge, US Geological Survey Water Resources Investigations Report 95-4218*, Washington, DC: US Government Printing Office, 37 pp.

Mosley, M.P. and McKerchar, A.I. 1992. Streamflow. In: Maidmont, D.R., ed., *Handbook of Hydrology*, New York: McGraw-Hill, pp. 8.1–8.39.

Muste, M., Fujita, I. and Hauet, A. 2008. Large-scale particle image velocimetry for measurements in riverine environments. *Water Resources Research* 44(4): W00D19.

Rantz, S.E., et al. 1982. *Measurement and Computation of Streamflow: Vol. 1. Measurement of Stage and Discharge, US Geological Survey Water Supply Paper 2175*, Washington, DC: US Government Printing Office, 631pp.

Richardson, E.V. and McQuivey, R.S. 1968. Measurement of turbulence in water. *Journal of the Hydraulics Division, American Society of Civil Engineers* 94: 411–430.

Schraub, F.A., Kline, S.J., Henry, J., Runstadler, P.W. and Littell, A. 1965. Use of hydrogen bubbles for quantitative determination of time dependent velocity fields in low speed water flows. *Journal of Basic Engineering* 87: 429–444.

Smith, J.D. 1978. Measurement of turbulence in ocean boundary layers. *Presented at the Working Conference on Current Measurement, Office of Ocean Engineering, National Oceanic and Atmospheric Administration, University of Delaware, Newark, DE, January* 11–13, 1978.

Smoot, G.F. and Novak, C.E. 1969. Measurement of discharge by the moving boat method. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A11.

Viessman, W., Knapp, J.W., Lewis, G.L. and Harbaugh, T.E. 1977. *Introduction to Hydrology*, New York: Harper and Row, 704pp.

Whiting, P.J. and Dietrich, W.E. 1991. Convective accelerations and boundary shear stress over a channel bar. *Water Resources Research* 27: 783–796.

Whiting, P.J., Stamm, J.F., Moog, D.B. and Orndorff, R.L. 1999. Sediment transporting flows in headwater streams. *Bulletin of the Geological Society of America* 111: 450–466.

Williams, A.J. 1996. Current measurement technology development in the '90s, a review. In: *Proceedings of the 1996 MTS/IEEE Oceans Conference,* Fort Lauderdale, FL, pp. 105–109.

Wilson, J.F. 1968. Fluorometric procedures for dye tracing. In: *Techniques of Water Resource Investigations of the United States Geological Survey*, Washington, DC: US Government Printing Office, Chapter A12.

CHAPTER 13
Measuring bed sediment

G. Mathias Kondolf[1] and Thomas E. Lisle[2]
[1]University of California, Berkeley, CA, USA
[2]McKinleyville, CA, USA

13.1 Introduction

Bed material is sampled for a range of purposes, including measurement of the framework size or the fine sediment content of spawning gravels, evaluation of substrate suitability for other habitat needs, as input for equations to calculate bed mobilization, bedload transport rates and likelihood of scour and as a measure of grain roughness in the channel. Particularly on gravel bed rivers, a variety of techniques have been used to sample bed material, ranging widely in effort and cost, mostly obtaining a gravel sample and passing it through a series of sieves to determine the proportions of various sizes or measuring particles under randomly located sample points on the bed. Although it may seem obvious, our principal message in this chapter is that the selection of sampling technique and analytical approach should be driven by the purpose of the study, i.e. the questions posed, the type of data needed to answer the question posed, the level of confidence needed in the result and consequently the requisite sample size. Many well-intentioned sampling programmes have produced data sets of ultimately questionable value because the purpose of the field data collection effort was not clearly thought out or sample size requirements were not recognized.

Theoretical and practical considerations for sampling gravel beds have been thoroughly reviewed in excellent works by Church *et al.* (1987) and Kellerhals and Bray (1971). Church *et al.* (1987) is a classic, dealing with fundamental issues of sample size, comparisons of different sampling methods and underlying study design issues. In this chapter, we review these considerations and specifically consider issues that arise in sampling for purposes such as assessing the quality of aquatic habitat or effects of upstream land use activities (Lisle and Eads 1991; Young *et al.* 1991). Sand and finer-grained sediments can be adequately analysed with relatively small samples, but gravels require large samples and thus pose greater challenges in sampling. A considerable literature in fluvial geomorphology concerns sampling of gravels, hence much of this chapter relates to sampling gravel bed rivers, sampling methods most appropriate for various objectives and advantages and disadvantages of various methods.

13.2 Attributes and reporting of sediment size distributions

Natural streambed sediments consist of a mixture of sizes, commonly ranging from clay (< 0.004 mm) to boulders (> 254 mm). If both gravel and clay are present in the mixture, particle size may range over five orders of magnitude. Many sediments (and sedimentary rocks) are characterized by larger particles that make up the structure of the deposit (the framework grains), with finer sediments filling the pore spaces between the framework grains (the matrix) (Carling and Reader 1982). Some sediments contain so much matrix that most framework grains are not touching and thus not carrying the weight of the deposit; these are termed 'matrix-supported' deposits (Williams *et al.* 1982). The threshold size between framework gravel and matrix sediment should be a function of the pore sizes in the framework. In a bimodal distribution, the distinction between framework and matrix may be straightforward. Otherwise, defining the upper limit of matrix sediment may be arbitrary, although matrix size distributions can be generated analytically from surface and subsurface size distributions (Lisle and Hilton 1999; Whiting and King 2003).

For each grain, three perpendicular axes can be identified: a long axis or *a axis*, a short axis or *c axis* and an intermediate axis or *b axis* (Krumbein 1941). Grain diameter is usually measured by the intermediate axis.

The range in particle size of natural sediments is continuous, but we customarily subdivide the range into size classes for standard terminology (e.g. sand, silt) and to yield sufficient classes for analysis (Pettijohn 1975). Because the range of sizes in natural sediments is so great, it cannot be effectively captured with a linear scale, so instead, a geometric scale is used.

The most commonly used size scale in fluvial geomorphology and engineering is the Wentworth scale, which defines size classes in millimetres and with intervals that increase by powers of 2 (Fig. 13.1). Common nomenclature of size classes (e.g. medium gravel at 8–16 mm) corresponds to the Wentworth scale. There is also a considerable literature (especially in engineering and biology) that has reported sizes in inches and many commonly used sieves are sized in fractions of inches. Fluvial

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

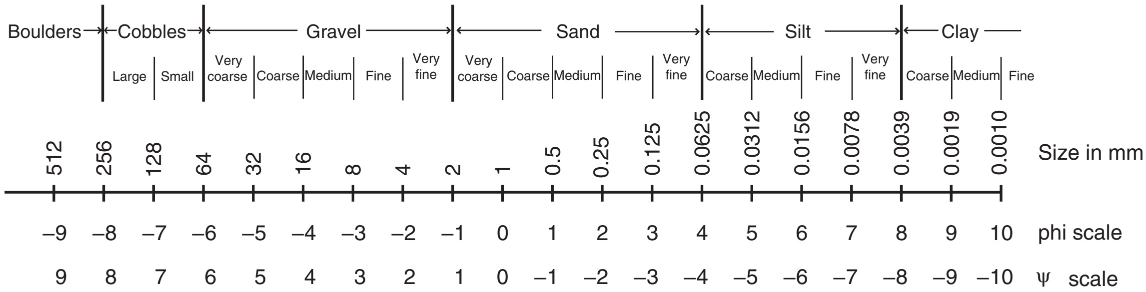


Figure 13.1 The Wentworth grain size scale, with equivalent *phi* values of Krumbein (1934). Source: Krumbein, 1934. Reproduced with the permission of Geoscience World.

gravels span such a wide range of grain sizes that their distributions are usually plotted either with log-transformed axes or the size data themselves are log-transformed to the so-called *phi* scale (Krumbein 1934; Inman 1952). The *phi* scale consists of size units corresponding to powers of two (in millimetres), but with the *phi* values increasing with decreasing size such that *phi* = 0 is equivalent to 1 mm, *phi* = 1 is equivalent to 0.5 mm, *phi* = 2 is equivalent to 0.25 mm, *phi* = −1 is equivalent to 2 mm, etc. However, to avoid the counterintuitive decrease in the scale value with increasing particle size, some geomorphologists have used the negative of the phi scale, termed the *psi* scale (Parker and Andrews 1985; Bunte and Abt 2001). The log-transformation to the *phi* scale provided a computational advantage when it was introduced in 1934, allowing computations to be easily made despite a wide range of grain sizes in natural sediments. However, this computational advantage is no longer meaningful with current computational capabilities. Because the *phi* and *psi* size classes are less readily comprehended than actual grain sizes values expressed in millimetres, we recommend reporting and plotting actual grain sizes (in millimetres), thereby increasing clarity of communication. There is simply no need for the added jargon of *phi* or *psi* values. With log-transformed scales, actual grain size values for different percentiles can be easily read from size distribution curves.

Presenting particle size distribution curves

Particle size distributions can be presented as histograms of the percentage of particle (or sample weight) occurring in each size class, as cumulative size distribution curves or as box-and-whisker plots. Unless the intervals between sieve sizes follows a geometric progression such as the Wentworth classes (and in many published studies the sieves did not), plotted non-cumulative particle size distributions can be misleading. One bar on the histogram may appear larger than the next only because the bar includes particles from a wider range of sizes. Therefore, the range of sizes present in a natural sediment is typically presented in cumulative size distribution curves (Fig. 13.2). Grain diameters corresponding to specific percentile values can be read directly from the curves plotted on a semilogarithmic scale (percentiles plotted on the *y*-axis, grain sizes plotted on a logarithmic scale on the *x*-axis) or by linear interpolation. The D_{16} is the size (in millimetres), at which 16% of the sample is finer, D_{25} the size at which 25% is finer, etc. Probably the most widely used percentile value is D_{50}, the median diameter.

Although these cumulative size distribution curves, if adequately sampled, can provide complete information on the range of sizes present in a given gravel, comparisons between gravels can be unwieldy and attributes of individual distribution obscured when too many similar distributions are presented together.

Size distributions can also be presented as modified box-and-whisker plots (Tukey 1977; Kondolf and Wolman 1993), which permit multiple distributions to be presented on the same graph without overlap (Fig. 13.3). In the box-and-whisker plots, the rectangle (box) encompasses the middle 50% of the sample, from the D_{25} to D_{75} values, termed the 'hinges'. The median diameter, D_{50}, is represented by a horizontal line through the box. Above and below the box are lines (whiskers) extending to the D_{90} and D_{10} values, a modification from the standard box-and-whisker plot of Tukey (1977), in which the whiskers extend out to extreme values. In the case of sediment size distributions, the range of sizes from gravel to clays is so great that it is not practical to plot whiskers to the extremes, so the D_{90} and D_{10} values can be used to capture the range of most of the size distribution.

Statistical descriptors

To facilitate comparison among size distributions, statistics are commonly drawn from the curves for comparison. For example, the median particle diameter, D_{50}, is commonly used in hydrology, geomorphology and engineering as a measure of central tendency of the distribution because it is easily read and unambiguously interpreted (Inman 1952; Vanoni 1975). Gravel size distributions tend to resemble log-normal, gamma or Weibull distributions rather than normal distributions (Kondolf and Adhikari 2000). Otto (1939) noted the resemblance of grain size distributions of sediments to the log-normal distribution and as a measure of central tendency, proposed the geometric mean in lieu of an arithmetic mean. The geometric mean is a measure of central tendency complementary to the median diameter and more influenced by extremes of the distribution. The geometric mean ($D_g = \Sigma f_i D_i$, where f_i is the fraction of the sample of size class i represented by its logarithmic mean

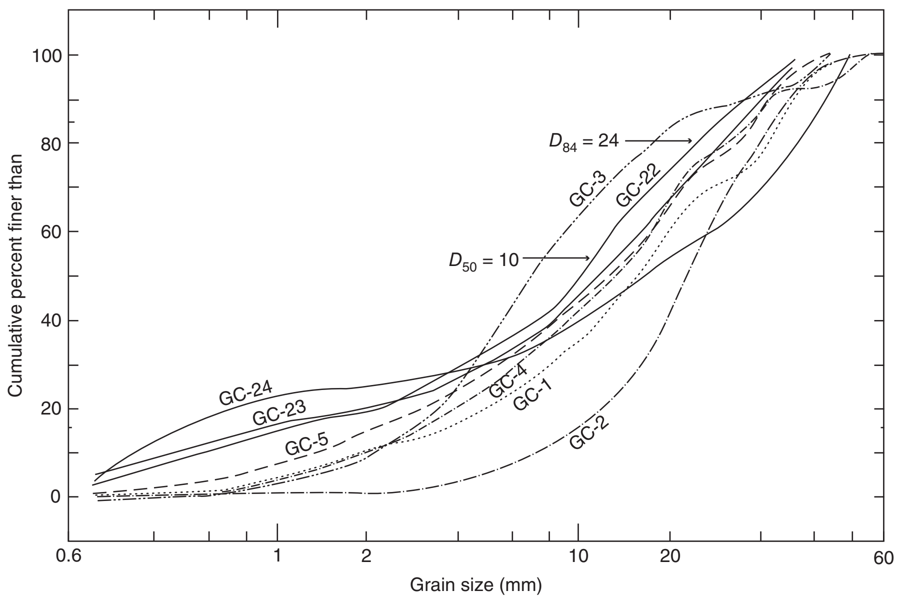


Figure 13.2 Illustrative cumulative size distribution curves for rainbow trout (*Oncorhynchus mykiss*) spawning gravels drawn from the case study in the mainstem Colorado River (solid lines, 3 samples) and Nankoweap Creek (dashed-dotted lines, 5 samples), a tributary downstream of Glen Canyon Dam. Size descriptors (D_{50}, D_{84}, etc) are read from the curve by reading the grain size corresponding to the indicated percentile. In the example shown, a redd gravel from Four-Mile Bar in the mainstem Colorado has a D_{84} of 24 mm and a D_{50} of 10 mm. Curves are identified by sample numbers (same as in Figure 13.3) (adapted from Kondolf 1988). Source: Kondolf, 2000. Reproduced with the permission of American Fisheries Society.

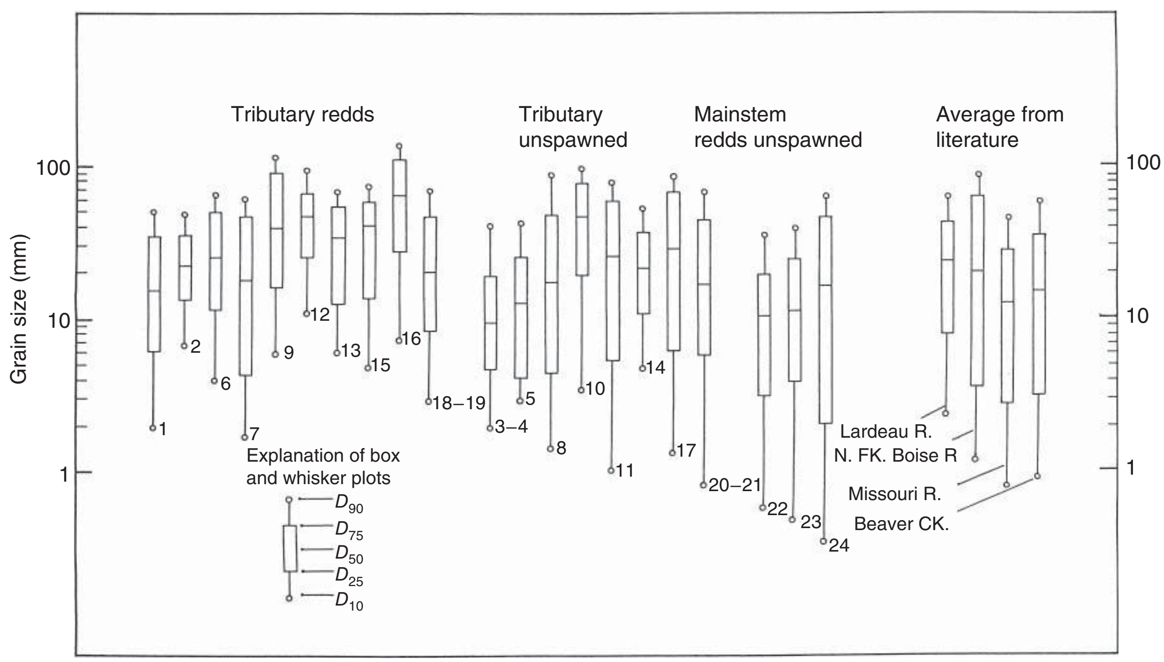


Figure 13.3 Box-and-whisker plots for rainbow trout (*Oncorhynchus mykiss*) spawning gravels from the case study in the Colorado River and tributaries downstream of Glen Canyon Dam and averages for other rainbow trout spawning gravels. For each sample, the rectangle (box) encompasses the middle 50% of the sample, from the D_{25} to D_{75} values, termed the 'hinges'. The median diameter, D_{50}, is represented by a horizontal line through the box. Above and below the box are lines (whiskers) extending to the D_{90} and D_{10} values. Numbers refer to samples in Kondolf *et al.* (1989). Box-and-whisker plots are easier to read than cumulative size distribution curves when multiple, similar distributions are plotted on the same graph. Source: Kondolf (2000), reproduced with permission of the American Fisheries Society.

D_i) can be calculated from the full size distribution using an algorithm in a spreadsheet. It can also be calculated by using the D_{16} and D_{84} sizes found graphically and assuming a normal distribution in which the 16th and 84th percentiles lie one standard deviation from the mean (Table 13.1). However, the assumption of a normal distribution in sediment size distributions is commonly inaccurate and the computational approach (e.g. Gary Parker's Morphodynamics Web Page: hydrolab.illinois.edu/people/parkerg/excel_files.htm) provides an alternative approach to calculating the geometric mean and geometric standard.

Other commonly reported attributes of size distributions are sorting and skewness. Sorting, or dispersion, refers to the degree of concentration or dispersion among the particles. Sorting in fluvial sediments is the process by which particles of a given size are concentrated. In geological parlance (as followed here), 'well sorted' means of similar size. In engineering usage, the same term may be used for a well-dispersed size distribution. In downstream reaches of larger river systems, currents may deposit bars composed entirely of gravel, other bars entirely of sand. These deposits would be considered 'well sorted' or having low dispersion. The Trask (1932) sorting coefficient is based on quartile values drawn from the size distribution and has been widely used in geological studies, but has been largely replaced by the geometric sorting coefficient, *sg*, of Otto (1939) (Table 13.1), based on the D_{16} and D_{84}. Both of these sorting coefficients increase with dispersion (and thus decrease with sorting). Sorting can also be quantified by the geometric standard deviation, expressed in *phi* units as $\sigma_g = [\Sigma f_i(D_\mathrm{g} - \mathrm{D}_i)^2]^{0.5}$, calculated from the full size distribution using a spreadsheet. All of these sorting parameters are equivalent, although graphical and computational methods use different attributes of the size distribution.

Skewness (*sk*) refers to the degree to which the distribution is skewed from a normal or lognormal distribution. Again, there are both quartile- and geometric-based skewness coefficients (Table 13.1). If plotted on an arithmetic scale, gravel size distributions tend to be positively skewed, which is to say that the coarse tails extend further than the fine or the mode is shifted towards the coarse end of the size distribution. However, when plotted on log-transformed scale, distributions tend to be negatively skewed, so the geometric mean diameters tend to be less than median diameters, as reflected in negative values of *sk* from a wide range of gravel size distributions (Kondolf and Wolman 1993). Kurtosis refers to the peakedness of the distribution and can be calculated from D_{10}, D_{25}, D_{75} and D_{90} (Kelley's quartile kurtosis in Krumbein and Pettijohn 1938) or D_{05}, D_{16}, D_{84} and D_{95} (Inman 1952) (Table 13.1).

Table 13.1 Size descriptors commonly drawn from sediment size distributions.

Measure of	Quartile-based descriptors	Descriptors based on D_{16}, D_{84}
Central tendency	Median D_{50}	Geometric mean (Otto 1939) $D_\mathrm{g} = [(D_{84})(D_{16})]^{0.5}$
Dispersion	Trask sorting coefficient (Trask 1932) $s_i = (D_{75}/D_{25})^{0.5}$	Geometric sorting coefficient (Otto 1939) $sg = [(D_{84})(D_{16})]^{0.5}$
Skewness	Quartile coefficient of skewness (Krumbein and Pettijohn 1938) $SK = [(D_{75}D_{25})/(D_{50})^2]^{0.5}$	Geometric skewness coefficient (Inman 1952) $sk = \log(D_\mathrm{g}/D_{50})/\log(sg)$
Peakedness	Kelley's quartile kurtosis (Krumbein and Pettijohn 1938) $KR = (D_{75} - D_{25})/2(D_{90} - D_{10})$	Geometric kurtosis (Inman 1952) $kr = \log[(D_{16}D_{95})/(D_{05}D_{84})]/\log(sg)$

Source: Kondolf and Wolman, 1993. Reproduce with permission of Elsevier.

Arguing that fluvial gravels were not log-normally distributed (and certainly many are not), Beschta (1982) suggested that use of measures derived by analogy with the log-normal distribution (D_{16}, D_{84}) were not valid and (with Lotspeich and Everest 1981 and Shirazi *et al.* 1981) proposed calculating moment measures as an alternative. The moment measures of traditional statistics are defined in terms of moments about the origin (for the first moment, the mean) and moments about the mean (for higher moments) of individual observations. However, sediment size distribution from bulk samples are obtained from a limited number of observations (sieves) and reported in percent weight, and result in open-ended curves that do not give the coarse and fine limits of the distribution. Because moment measures depend on the entire size distribution, this limits their application to sediment grain size distributions (Inman 1952).

Accordingly, the usual practice is to draw values from the cumulative size distribution and compute measures from them (Table 13.1) (Vanoni 1975). The Trask (1932) measures are based on the central 50% of the distribution only, whereas the geometric mean, sorting index, skewness and kurtosis measures of Otto (1939) and Inman (1952) are based on percentile values that encompass more of the distribution, and so provide better measures of many attributes of the size distributions. As a measure of central tendency, the median size, D_{50}, is arguably the best, as it is least affected by the tails.

The question of whether fluvial gravels fit a log-normal or 'Rosin' distribution has been debated in a number of publications prior to the recognition that the Rosin distribution is actually the same as the Weibull distribution, well studied in the probability and statistics literature (Kondolf and Adhikari 2000). The log-normal, Weibull and gamma distributions are so similar that it can be difficult to distinguish which is the best fit to a given gravel distribution. Earlier studies plotted cumulative particle size curves onto on a log-normal and Weibull-transformed scales and determined the r^2 values of the fit as a basis for determining which distribution best fit the data (e.g. Beschta 1982; Ibbeken 1983). However, because these distributions are so similar, high r^2 values are obtained even when an ideal Weibull ('Rosin') distribution is plotted on a log-normally transformed axis (Kondolf and Adhikari 2000). More fundamentally, the relation between the underlying mathematical properties of the different distributions and the physical mechanisms giving rise to them is not clear.

The choice of which descriptor to use depends largely on the purpose of the study. For example, D_{50} seems useful in transport relations (e.g. Shields stress) because its entrainment threshold appears to be constant, for a given range of slope, for varying degrees of sorting. D_g was used by Parker (1990) for the same purpose. Note that although D_g is usually close to D_{50}, the difference between these two measures increases with increasing skewness. As discussed in Section 13.11, both D_{50} and percentage finer than 1 mm are useful descriptors to assess the quality of salmonid spawning gravels.

13.3 Particle shape and roundness

Grain shape varies widely, reflecting properties of the source rock and subsequent weathering and abrasion. Zingg (1935) recognized four basic shape classes defined by ratios of the principal axes: oblate, equant, bladed and prolate (Fig. 13.4a). Grain shape can strongly influence the sediment transport characteristics of particles. The influence of shape on the settling velocity of particles in water is quantified by the Corey shape factor ($CSF = c/(ab)^{0.5}$, where a, b and c are the diameters of the longest, intermediate and shortest axis, respectively) (Dietrich 1982). Grain roundness is a distinct concept from shape: it is the degree to which sharp edges and corners of rock fragments have been removed by weathering and abrasion. Particles falling into the shape classes described above can exist in angular (unrounded) form or they can be transformed into rounded condition. Rounding has conventionally been addressed visually using charts such as shown in Fig. 13.4b (see also Bunte and Abt 2001). More recently, digital image analysis has been used to compute particle roundness (Roussillon *et al.* 2009) based on discrete geometry (Fig. 13.5). Whereas previous work used Fourier transformation (e.g. Diepenbroek *et al.* 1992), the discrete geometry approach allows the implementation of Wadell's original index (Wadell 1932), known to be more accurate but more time consuming to implement (Pissart *et al.* 1998). Wadell defined his roundness index as follows:

$$rW = \frac{1}{kR}\sum_{i=1}^{k} r_i \tag{13.1}$$

where r_i is the radius of curvature that is smaller than or equal to the radius of curvature R of the largest inscribed disk at a pixel on the boundary of the pebble silhouette and k is the number of such radii.

Other roundness measures include that proposed by Drevin and Vincent (2002) or the ratio between the perimeter of the silhouette and of the best approximating ellipse (Roussillon

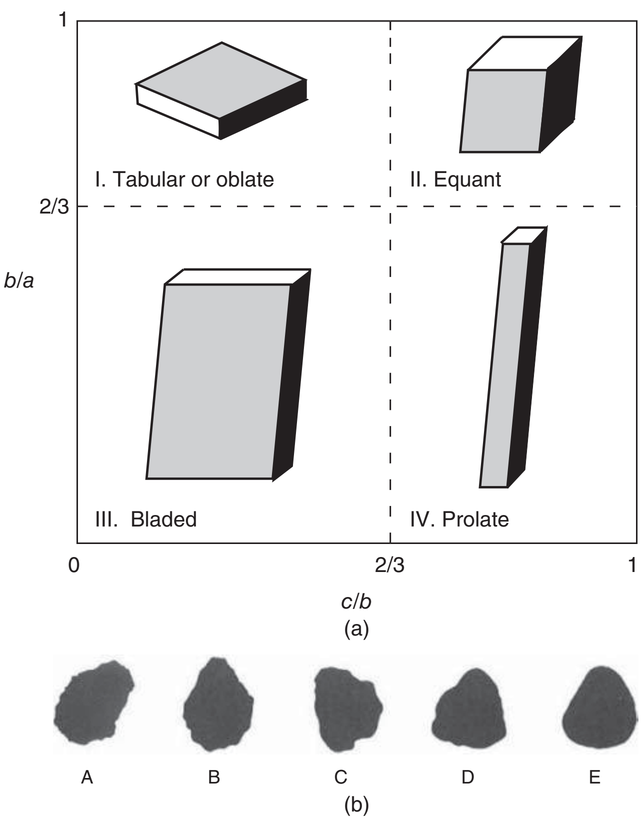


Figure 13.4 Grain shape and roundness. (a) Zingg shape classes, plotted by ratios of principal axes. (b) Roundness classes: A, angular; B, subangular; C, sub-rounded; D, rounded; E, well-rounded. Based on Zingg (1935) and Krumbein (1941).

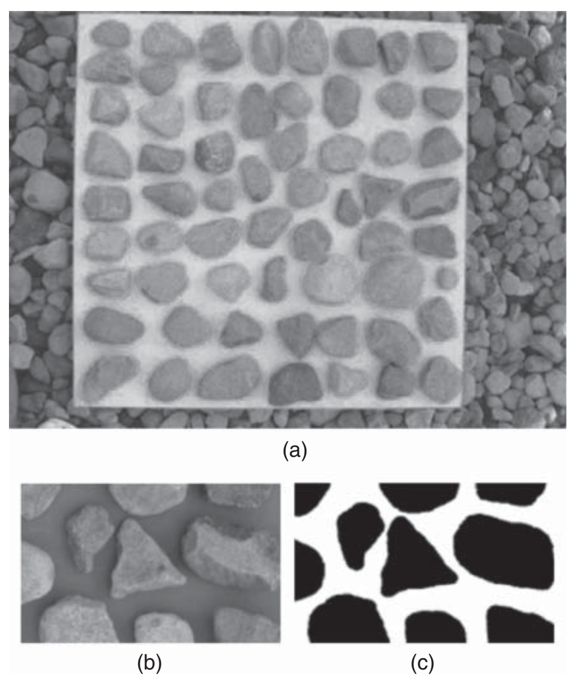


Figure 13.5 (a) An image of sample pebbles with boundaries extracted in white. (b) Extraction is performed by contour tracking in binary image. (c) Computed with clustering methods applied to the original colour image. Source: Roussillon *et al.* (2009), reproduced with permission of Elsevier.

et al. 2009). Wadell's index calculated from imagery is highly correlated (92%) with the roundness classes of Krumbein's chart (1941), which can be used as a ground-truth, as can the ratio of perimeters (see fig. 15.5, p. 349, in Bertoldi *et al.* 2012). Roundness can serve as an identifying characteristic of particles and provides an indication of the history of transport of the particle from the source rock. However, there is no universal relation between particle roundness and distance or duration of fluvial transport, because rocks can round as they weather in place in floodplain and terrace deposits and because rounded clasts can be eroded from conglomerates and recycled.

13.4 Surface versus subsurface layers in gravel bed rivers

The surface layer of a gravel bed is commonly coarser than the underlying, subsurface layers (Fig. 13.6). The size distribution of the subsurface gravel is commonly similar to or slightly coarser than that of the transported bedload (Parker and Klingeman 1982; Lisle 1995). The framework grains of the surface are generally not larger than those of the underlying sediment, but the surface layer is typically deficient in the finer fractions of the distribution. In part, this can be explained by selective transport of finer grains exposed on the surface at flows too low to mobilize the entire bed. The paucity of interstitial fine sediment in the surface layer implies that while framework size can be estimated by sampling the surface layer, matrix assessment requires subsurface sampling.

Some coarse surface layers are active features in that they persist (or re-establish) despite frequent mobilization of the bed and are common features of gravel beds with active sediment transport. These surface layers were termed *pavements* by Parker and Klingeman (1982), as distinct from inactive coarse surface layers that result from the progressive winnowing of finer fractions from the surface layer, as might be encountered below a dam, which they termed *armour layers*. Gomez (1984) proposed similar distinctions, but argued that the terms should be used in an opposite fashion: armour for active surface layer, pavement for inactive surface layer, by analogy with the fact that armour (like medieval armour) can be removed, whereas pavement (like a road surface) cannot. In recent discourses, we have encountered terms such as 'active armour', and many avoid the issue by simply referring to a 'coarse surface layer', an approach with the virtue of avoiding inferences of degree of mobility when there are no direct observations of such.

Figure 13.6 Surface and subsurface texture in gravel bar in the Garcia River, California. Photograph by G.M. Kondolf, April 1992.

Differences between surface and subsurface grain sizes can help to evaluate variations in bed mobility associated with sediment supply (Dietrich *et al.* 1989; Buffington and Montgomery 1999b; Lisle *et al.* 2000). Increases in the bed material load result in an increase in transport intensity at a given flow magnitude. This is accommodated by a decrease in surface particle size approaching that of the load or subsurface, but the adjustment is mediated by the magnitude of boundary shear stress exerted on the bed. The references cited detail the methods used to measure indices that are based on this adjustment. The index q^* is the ratio of bedload transport predicted from the particle size distribution of the load (commonly represented by the subsurface material) to that predicted from the size distribution of the bed surface, given a reference boundary shear stress (Dietrich *et al.* 1989). As armouring decreases, values of q^* generally increase towards 1, with the value 1 signifying the absence of armouring. Shields stress is more commonly used to scale bed mobility (Yalin 1977). It is the ratio of the forces of traction and gravity acting on a representative size fraction (usually D_{50}) in a river bed or more precisely $T^* = T/RD_{50}$, where T^* is boundary shear stress and R is the submerged specific gravity of the sediment.

In fish habitat studies, the distinction between surface and subsurface populations has not always been acknowledged. If habitat for fry or aquatic insects is the concern, then the surface population should be sampled. If intragravel condition for incubating salmonid embryos is the concern, then the subsurface population should be sampled (Kondolf 2000). If the size of framework gravel selected for spawning by a species is the concern, then either surface or subsurface sampling will yield reasonable results, unless the bed is so armoured that the two populations are extremely different. In completed salmon redds (nests containing incubating salmon eggs), the surface and subsurface populations have already been mixed, although coarser gravels may have been concentrated at the base of the redd as lag deposits in the egg pocket.

13.5 Sampling sand and finer grained sediment

Sampling sand and finer grained sediments is relatively straightforward, provided that there is access to sample sites. For these sediments, one litre is generally an adequate sample size. Sand can be sampled from the exposed bed or bars with a shovel or trowel, from underwater sites (and under high flow conditions) with various bed material samplers including drag bucket, grab bucket and vertical pipe type samplers (see Vanoni 1975, pp. 334–337, for a description). Samples of sand are typically dried and passed through a series of sieves with progressively finer mesh sizes, yielding particle size distributions. Large samples are commonly split into smaller subsamples for analysis. For silt and clay, grain sizes must be determined by sedimentation apparatus, such as settling tubes, elutriation (washing lighter particles off, leaving heavier particles behind) and centrifuge separation (Vanoni 1975).

The stratigraphy of channel deposits (the vertical and horizontal arrangement of sedimentary layers) and their grain size and lithology can yield information on depositional history and channel dynamics (see Chapter 2).

13.6 Sampling and describing the surface of gravel beds

Sampling methods can generally be divided into those that sample the surface and those that sample the subsurface. Surface methods include facies mapping, visual estimates, pebble counts and photographic methods. Subsurface methods include shovel or backhoe samples (from exposed bars), core samples collected with cylindrical samplers driven into the bed, core samples obtained by freezing interstitial water around a probe in the bed, and those obtained by dredge.

Facies mapping

The term *facies* refers to a mappable area of the streambed that can be delineated/mapped based on particle size, representing distinct local depositional environments (Pettijohn 1975). A less sedimentological synonym is *patch* (Seal and Paola 1995). Both refer to areas (with distinct grain size) whose long dimension is on the scale of half the channel width or more. A facies (or patch) may consist of a mixture of poorly sorted grains of many sizes, but it should be consistently so over the entire patch. Where more than one facies exist in a reach, the facies can be visually distinguished (on exposed bars or in clear water) and mapped (Lisle and Madej 1992; Seal and Paola 1995; Buffington and Montgomery 1999a). This approach can be used not only to distinguish deposits of sand from gravel, but also to distinguish different sand–gravel mixtures, as field trials have shown observers are capable of doing (Shirazi and Seim 1981).

Facies maps can be extremely useful tools as descriptors of current conditions (from which relative proportions of different sized units can be measured), as baseline data against which to measure future change or as a basis for comparing sediment conditions among channels. Their primary advantage is that they capture reach-wide variations in surface size and are not subject to the greater variability commonly encountered at smaller channel scales (e.g. pebble counts at cross-sections), which can be affected by changes in channel morphology. Moreover, facies maps coupled with channel morphology are effective for interpreting channel processes. For example, Wolman and Schick (1967) used facies maps to document the influence of construction-derived sediment on Oregon Branch north of Baltimore, Maryland (Fig. 13.7). Beaverdam Run, a tributary that was unaffected by the recent construction, consisted of cobble riffles and pools, with small deposits of fine humic sediments in convex portions of normal bends. By contrast, Oregon Branch below the construction area contained silt and sand deposits throughout, some as thick as 60 cm (Wolman and Schick 1967).

The pebble count or grid sampling

The pebble count (Wolman 1954) is a sampling of approximately 100 grains (stones) on the river bed (or gravel bar), on a grid or line. As an alternative to a grid, sampling points can be selected by picking grains encountered in front of the observer's boots at regular intervals as they proceed across the bed. Provided that the selection of stones is truly random, the two methods yield equivalent results. The pebble count can be conducted on an exposed gravel bar, by wading in shallow water or, in greater water depths, by diving (Klingeman and Emmett 1982). The stone measured at each sample point is selected randomly by dropping the finger to the bed with eyes closed or averted to avoid bias towards larger particles. When sampling under water, there may be a bias towards larger particles because fingers are displaced by the current from a truly vertical descent (Marcus *et al.* 1995).

Ruler versus template

The intermediate axes of the stones are measured either with a pocket ruler and recorded within predetermined size classes or passed through a template in which squares have been cut in the sizes of the grain size classes, analogous to sieve openings (Hey and Thorne 1983). For well-rounded stones, the two methods are virtually equivalent. However, flattened, elongated clasts can pass through square template openings with a diameter smaller than their actual *b*-axis length, because they can orient diagonally through the openings, resulting in smaller measured sizes for the same stones (Church *et al.* 1987). Thus, although they may not measure the true *b* axis, template measurements are less prone to observer error by inexperienced field personnel in identifying the *b* axis to measure, and are more comparable to sieve measurements. Empirical bed load transport formulae are scaled in part by sieve size, as distinct from the true *b*-axis size, so in this respect a flaw of the template may be viewed as a virtue because it is more consistent with sieves. In any event, when reporting methods for pebble counts, the method used should be clearly stated and the shape of particles can be described or quantified using standard sedimentological particle shape factors.

Whether sampled by grid or pacing, the sampling point should serve only as the starting point for the finger's blind descent to the bed. If the finger touches more than one stone simultaneously (perhaps lodging between two stones), bias is

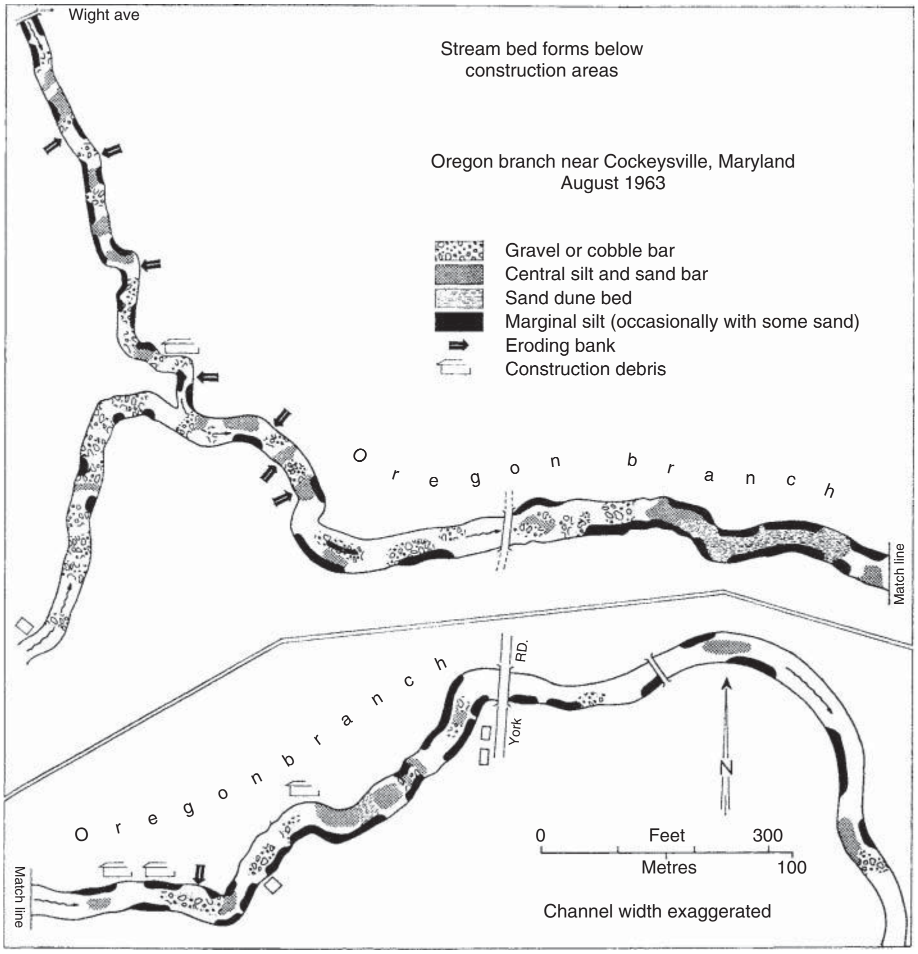


Figure 13.7 Facies map for Oregon Branch, near Cockeysville, Maryland, showing extent and pattern of silt and sand deposits downstream of rapidly eroding construction area, in contrast to the limited fine sediment deposits in the channel of Beaverdam Run, a tributary unaffected by recent construction activity. Source: Wolman and Schick, 1997. Reproduced with the permission of AGU.

introduced unless the operator consistently uses one point on the finger (such as the right corner of the fingernail) as the sample point. At the outset of the sampling programme, decisions should be made about how to classify situations such as a thin layer of sand over a larger stone, etc. In cobble and boulder beds, templates cannot be used because particles are too large and frequently embedded and the pacing technique may not be truly random because the operator's instinct to preserve their shins will influence their pacing. We suspect that one common source of error is probably the failure to close fully or avert the eyes as the finger descends to the bed and the resultant attraction to larger (more visually discernible) particles. The *b* axis should be measured perpendicular to the long (*a*) axis at its widest point. Using a template should minimize differences in measurement of the *b* axis among inexperienced operators.

Size intervals

As the stones are sampled, their sieve (or *b*-axis) sizes are recorded in grain size classes (in millimetres) that increase by powers of $2^{0.5}$. The grains are recorded on the row identified by the lower end of the size range, by analogy with sediments collected on sieves. A 2 mm sieve collects all grains smaller than the next largest sieve size (4 mm for sieves following the *phi* scale), but larger than 2 mm. Similarly, in a pebble count, a stone with intermediate axis of 52 mm would fall in the 45 mm class because it is smaller than 64 mm but larger than 45 mm. When the finger encounters sediment finer than 4 mm (or 8 mm), it is recorded as '<4 mm' (or '<8 mm'). The results are recorded in the field book as tick marks, yielding a histogram in the field book (Fig. 13.8). A 52 mm stone would be recorded by a tick on the row labelled '45 mm'; a 44 mm stone would be recorded by a tick on the row labelled '32 mm'. The total in each class divided by the total sample number yields cumulative percentages for each size class (easily calculated on a spreadsheet, hand calculator, slide rule or paper).

These are the cumulative percentages *finer* than the next largest class size (i.e. the class size immediately above it on the table). In the example shown in Fig. 13.9, adding the percentages from the

Rush Creek at IFIM site 8/28/87

XS 36 GMK, SSC

AVK

Stratified populations A,B,C (sand)

Population 'A'					
Size class (mm)	Number of rocks	Number of rocks	%	Cum %	Finer than (mm)
256	II	2	1.6	100	360
180	IIIE	5	4	98.4	256
128	IIII IIEE IIII E	16	12.8	94.4	180
90	IIII IIII EIII IEII IEEE IEII	30	24	81.6	128
64	IIII IIII EIII IIII IEEI EIII	30	24	57.6	90
45	IIII IIII IIII IIII	19	15.2	33.6	64
32	IIII I	6	4.8	18.4	45
22.6	IIII II	7	5.6	13.6	32
16	IIII	4	3.2	8.0	22.6
11.3	II	2	1.6	4.8	16
8	I	1	0.8	3.2	11.3
< 8	III	3	2.4	2.4	8
		$\Sigma = 125$			
		$\Sigma > 8$ mm = 122			

Total embedded = 14, all > 64 mm
= 11% of all rocks

Figure 13.8 Field notes from a pebble count on Rush Creek, California, showing field-generated histogram. Note sample points on embedded rocks designated by *E*s instead of regular tick marks. Source: Kondolf, 1997. Reproduced with the permission of Wiley.

bottom up yields 33.6% on the 45 mm row. This is the cumulative percentage finer than 64 mm. To read the cumulative percentages easily, the sizes can be repeated in a column to the right of the cumulative percentages, with the sizes shifted downward by one row so the adjacent columns can be easily read as 'Cumulative percent finer than *size*'.

In some river beds, stones may be interlocked in the gravel matrix and difficult to remove. This is especially a problem with larger stones in armoured cobble beds or with boulders. In such cases, it may be impractical to remove every stone for a complete inspection and accurate measurement, so the observer can partially excavate the stone and estimate the size with a ruler. With the large sizes, the particle size classes are so widely separated in size (arithmetically) that one can be fairly certain of the size class of most such embedded particles unless they fall near the boundary between classes. In the field notes, such an embedded particle can be recorded with an 'E' (instead of a tick mark) in the appropriate size class (Fig. 13.9). The percentage of embedded stones in the sample can be calculated, providing a rough measure of the degree to which bed particles are interlocked, and also a measure of this source of error in sampling (Kondolf 1997). The pebble count can be adapted to yield other useful information, such as lithologies of different particles sampled. Instead of ticks, for each pebble measured a letter designating lithology can be recorded, such as 'A' for andesite, 'S' for sandstone, etc., depending on the lithologies encountered. Particles can also be recorded as competent or friable, based on their response to a standardized blow from a rock hammer.

Sampling by facies

The pebble count is conducted over a patch of gravel comprising a single facies or population, 'a zone or area considered homogeneous' (Dunne and Leopold 1978, p. 666). If only one facies can be distinguished in a reach, the grain size distribution can be applied to the entire reach. For large homogeneous areas, such as gravel bars on large rivers, which can be tens or hundreds of metres long, the pebble count can be conducted over part of the bed and its result applied to the entire homogeneous feature, or the pebble count can be conducted such that its sampled particles are drawn from over the entire feature, with the distance between sampled pebbles commensurate with the size of the area sampled. The first approach is based on the identification of the entire feature as 'homogeneous' and the corollary that a subsample should be representative of the entire population. The second

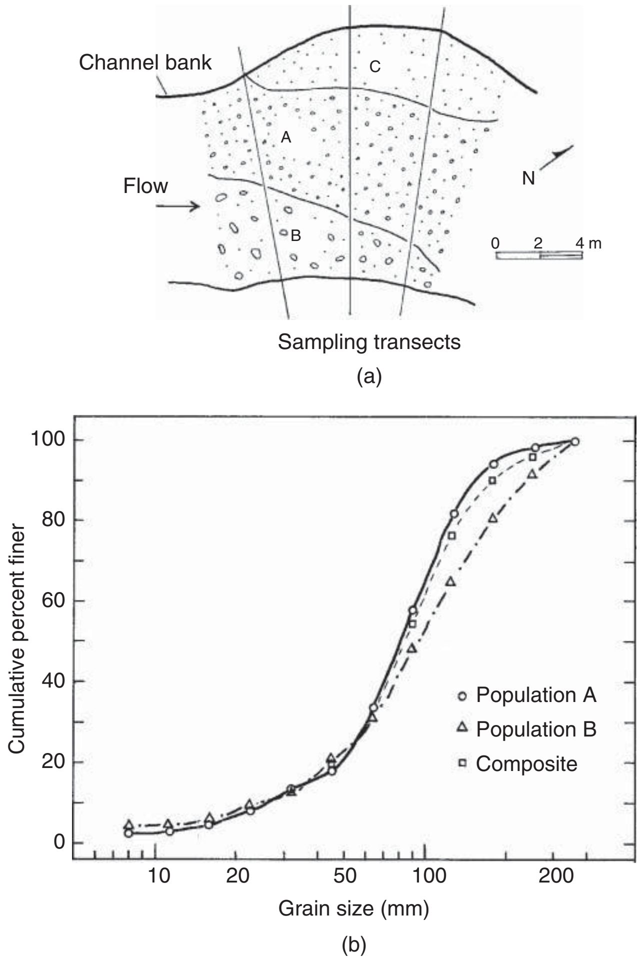


Figure 13.9 Bed material at a sample site on Rush Creek, California. (a) Facies map of distinct gravel–sand mixtures and sand deposits identified on the bed. The middle cross-section shown was a sampling transect used in an aquatic habitat study (Kondolf and Li 1992). The adjacent two cross-sections were established to provide enough sampling points for grid-based pebble counts. (b) Grain size distributions for facies A and B and composite distribution for the two based on their respective bed areas in the study reach. Source: Kondolf, 1997. Reproduced with the permission of Wiley.

approach defines the sampling universe (facies) and attempts to give each individual (pebble) an equal probability of being sampled. For a given facies, the two approaches should be equivalent.

In the case where two or more distinct facies exist in the reach to be characterized, the different facies can be mapped and measured and separate pebble counts conducted on each. This approach can be used not only to distinguish sand from gravel, but also to distinguish different gravel-sand mixtures that may be present in the channel. Field trials described in the literature indicate that observers are capable of visually distinguishing distinct facies (Shirazi and Seim 1981) and in fact that visual identification tended to overestimate differences in grain size among different facies of gravel–sand mixtures (Kondolf and Li 1992). If a composite grain size distribution for the entire reach is desired, the proportions of the bed occupied by each distinct bed material facies can be measured, pebble counts conducted on each and a weighted average grain size distribution computed (Fig. 13.9).

To address the pebble count's inability to sample particles smaller than about 4 mm, Fripp and Diplas (1993) and Petrie and Diplas (2000) proposed a hybrid technique in which the pebble count sample is truncated around 10 mm and the fine tail is defined by a completely different technique, areal sampling by adhesion to clay and the two samples are merged.

Visual estimates

Visual estimates, often grandly termed 'ocular assessments', involve estimating, by eye alone, the sizes of substrate particles or estimating percentages in different broad size classes. Casual descriptions of bed material size such as '2 inch gravel' are, in effect, visual estimates. Visual estimations of grain size have been used more formally and systematically by fisheries biologists and are the basis for many of the published descriptions of gravel sizes used for salmonid spawning, typically reported as a range of sizes preferred by the studied species, such as '1 to 4 inch gravel' (e.g. Greeley 1932; Hazzard 1932; Cope 1957; Hunter 1973). Visual estimates are the basis of the substrate code in the widely used instream flow incremental methodology (IFIM) (Bovee 1982), with a scale from 1 to 8 in ascending order of coarseness: plant detritus, clay, silt, sand, gravel, cobble, boulder and bedrock. Intermediate-sized substrates are reported using decimals, such as 5.2 to designate gravel with 20% cobble.

Despite the widespread use of visual estimates, we are unaware of any systematic studies demonstrating that these subjective estimates of percentages of various size classes in the bed are reproducible among different investigators. Moreover, even if these estimates are accurate, the results are usually reported in the form of a range of sizes (e.g. 1 to 3 inch gravel), 'dominant' and 'subdominant' size class or as percentages of classes such as '80% cobble, 10% sand and 10% silt'. Hence these estimates are not readily compared with sediment sizes reported in the engineering and geomorphic literature, in which statistics are drawn from standard size distributions.

Mapping the parts of the bed occupied by distinct facies and conducting adequately sized pebble counts on each facies is an approach that can be used to provide reproducible grain size information for aquatic habitat studies, in lieu of visual estimates at many points. For a bed with numerous facies present, the pebble count approach will be more time consuming than visual estimate at each point, but for a bed with only one facies, the pebble count approach is faster than visual estimates at many points and offers the advantage of yielding data consistent with those reported in the engineering and geomorphic literature (Kondolf and Li 1992).

Photographic grid methods

Image-based approaches to quantifying surface particle size have evolved significantly since the early efforts to analyse grain size distribution from photographs (e.g. Kellerhals and Bray 1971; Burns 1978; Adams 1979; Ibbeken and Schleyer 1986). Church *et al.* (1987) highlighted the issues of bias due to particle hiding and imbrication angle. Recent efforts using remote sensing and image processing analysis to estimate grain size on exposed beds have involved two main approaches. The first is based on geostatistical techniques applied to the textural properties of images. The texture of an image is calculated by assessing the differences in brightness values between neighbouring pixels using, for example, semi-variance, autocorrelation, entropy, contrast (Carbonneau *et al.* 2005; Verdu *et al.* 2005; Lejot *et al.* 2011; Black *et al.* 2014) or spectral analysis (Rubin 2004; Warrick *et al.* 2009; Buscombe *et al.* 2010). Using regressions, texture variables can be used for predicting grain size percentiles, from D_{10} to D_{90}. Such procedures have been applied in sand deposits as well as gravel, on exposed and submerged areas and to images acquired from low-elevation platforms such as helicopters, drones or balloons for mapping a given percentile (e.g. D_{50}) on an area (Verdu *et al.* 2005; Lejot *et al.* 2011) (Fig. 13.10) or along a long river reach, such as the longitudinal plot of median grain size (based on 4047 points) along 80 km of the Sainte Marguerite River (Canada) (Fig. 13.11) (Carbonneau *et al.* 2005). Aerial photosieving can serve to calibrate grain size maps, thereby reducing field effort (Dugdale *et al.* 2010).

The second main approach to using remote sensing and image analysis to estimate surface grain size uses image processing to identify the boundaries of individual grains in the image, building upon prior protocols. Following the work of Graham *et al.* (2005a,2005b, 2010, 2012), this approach has produced automatic algorithms which are now available online, such as Gravelometer (http://www.sedimetrics.com/) or BASEGRAIN (http://www.basement.ethz.ch/services/Tools/basegrain). Previously applied only to exposed bars, such as occur along 50 km of the Ain River (Rollet *et al.* 2014), new photo techniques can characterize grain size on channel beds too deep to show any textural pattern linked to grain size on aerial images (Pégot-Augier 2012). The main issues with this approach are

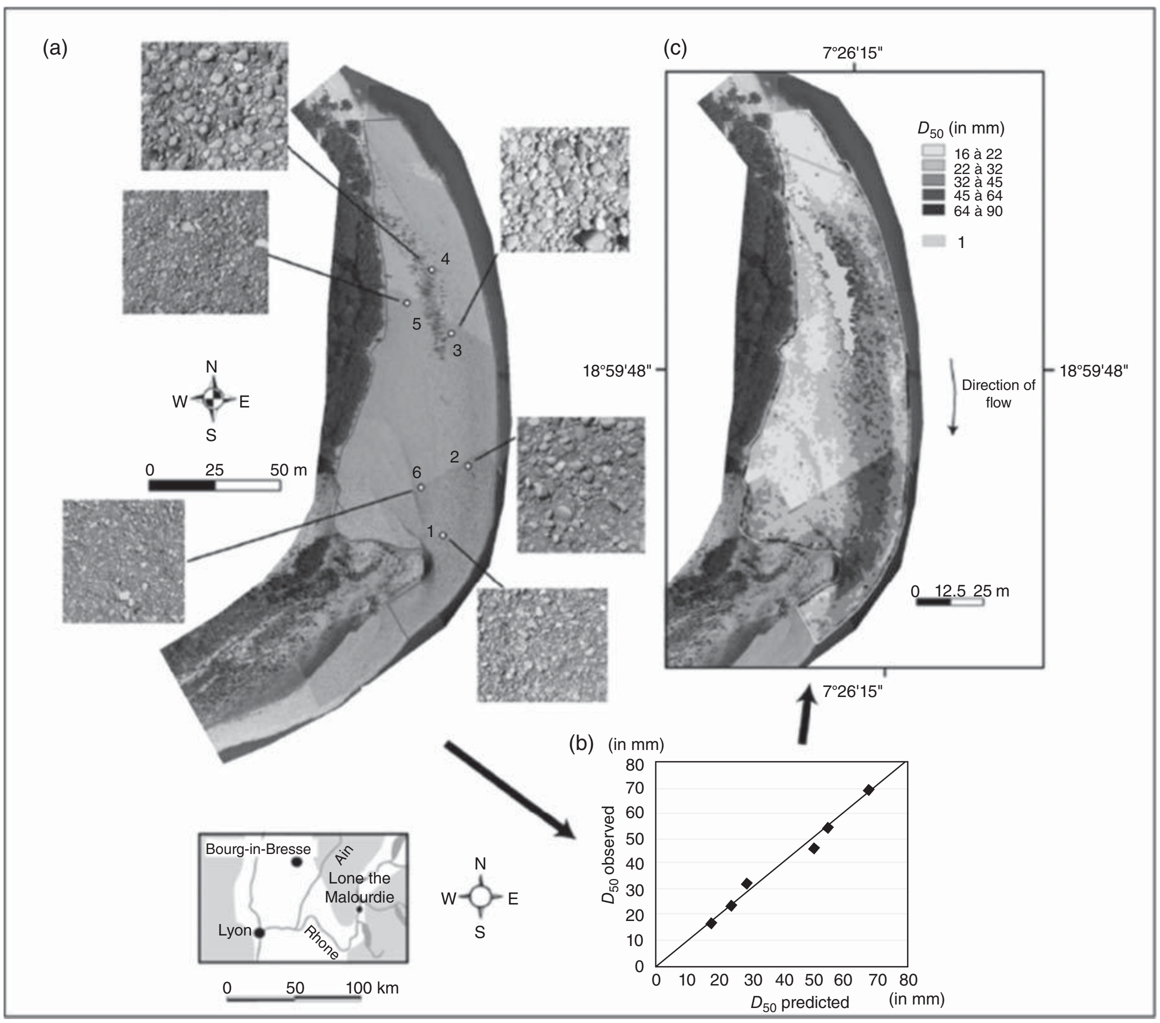


Figure 13.10 Grain-size mapping of Gevrieux gravel bar based on drone images with very high (3.4 cm) resolution (18 June 2004) (a), grain-size model (b) and mapping of predicted D_{50} by size class (c). Source: Lejot *et al*, 2011. Reproduced with the permission of Schweizerbart.

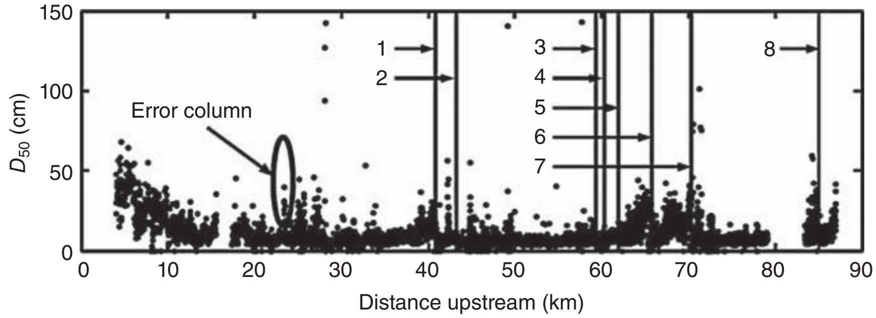


Figure 13.11 Long profile of median grain size on the Sainte-Marguerite River, Canada, showing link cutoff points (vertical lines), numbered 1–8 as determined by Davey and Lapointe (unpublished report, 2004) and an example of an 'error column' structure caused by glare at the water surface. Source: Charbonneau *et al*, 2005. Reproduced with the permission of AGU.

related to the fact that particles are overlapped so that the *b* axis measured is the *b* axis viewed and field calibration must be applied to take into account particle overlay or shape and sometimes petrography effects which complicate radiometric conditions and algorithm efficiency to detect particle boundaries. Specific protocols of image acquisition are also often necessary to control light conditions. Finally, these approaches apply only to cobbles and gravels, not sand, but yield the entire distribution of size classes, not only certain percentiles as the previous textural methods do. Using data sets collected from seven gravel bed rivers, Graham *et al.* (2010) assessed the minimum areas required to obtain representative samples, effects of lower end truncation on grain size percentiles, effects of river bed structure such as imbrication and hiding and potential benefits of using individual particle measurements rather than the number (or mass) of particles per size class to calculate percentiles. Because the different sampling and analysis techniques do not produce grain-size distributions that are directly comparable, Graham *et al.* (2012) explored the appropriate conversions between different types of surface grain-size sampling methods (e.g. errors associated with the use of empirically and theoretically derived conversion factors for image-based area-to-grid conversions, for area-to-grid and grid-to-area conversions, and for conversion between weight- and number-based samples) (Fig. 13.12).

When measuring roundness using the algorithm developed by Roussillon *et al.* (2009), the individual size of particles (a and b axes) can also be measured. To assess roundness, the grain must be fully in view, which requires that the samples be collected and the grains arranged on a board before being photographed.

13.7 Subsurface sampling methods

Bulk core sampling

Bulk core sampling involves directly removing a sample from the bed, usually within a predetermined area and down to a predetermined depth. Gravel exposed on a bar can be easily sampled by shovel or, even better for adequate sample size, a backhoe. In flowing water, bulk samples are commonly obtained by driving a cylindrical core sampler into the bed and removing (by hand) the material within. Geomorphologists have used bottomless 50 cm oil drums in various forms to obtain sufficiently large samples, such as the 140–240 kg samples collected by Wilcock *et al.* (1996b), the 50 cm 'cookie-cutter' drum sampler (Klingeman and Emmett 1982) and the 46 cm 'barrel' sampler (Milhous *et al.* 1995) (Fig. 13.13a). When removing the gravel from drum samplers, it is possible to remove the surface layer first and analyse it separately, often worthwhile given the difference between surface and subsurface layers.

A variant of the barrel sampler used in fisheries studies is the FRI or McNeil sampler, constructed from a 50 cm drum with a 15–30 cm diameter pipe welded on the bottom (Fig. 13.13b). The smaller pipe is worked into the bed, the gravel removed by hand and the muddy water within the sampler retained to permit suspended fine sediments to be sampled (McNeil and Ahnell 1964). The small pipe reduces the sample size, potentially to less than the minimum required to sample adequately the grain sizes present, and in gravels that include particles coarser than 50 mm the pipe edge may hit a large rock and cannot continue downwards unless the rock is moved out of the way and either included in the sample or discarded.

Freeze-core sampling

Freeze-core sampling involves driving steel probes into the bed, discharging a cooling agent (such as liquid carbon dioxide or nitrogen) into the probes to freeze the interstitial water adjacent to the probe and withdrawing the probes (with gravel samples frozen to them) from the bed with a tripod-mounted winch. The first versions of the method used a single probe, later versions used three probes (Everest *et al.* 1980). The method was developed largely to obtain gravel samples that preserved vertical stratification of the sediments, especially with respect to the vertical infiltration of fine sediments into salmon redds. However, laboratory experiments have shown that driving

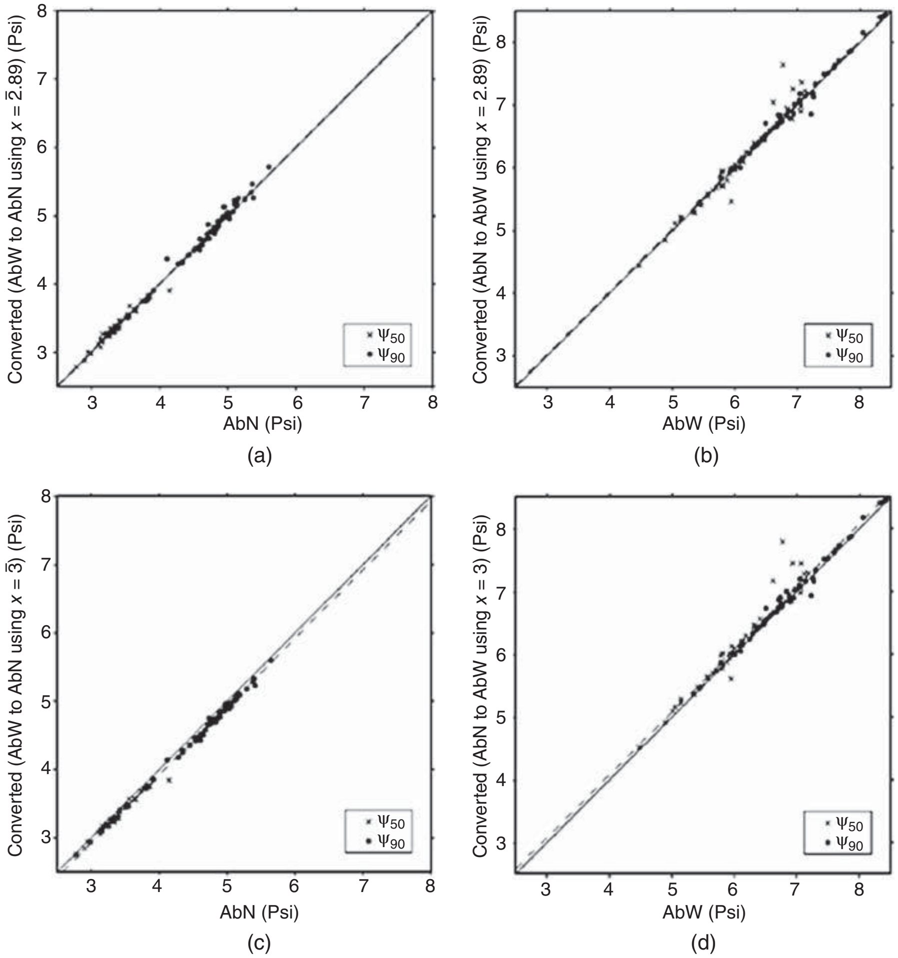


Figure 13.12 Covariant plot of the Ψ_{50} and Ψ_{90} for validation conversion factors according to sampling methods based on three field sites (Ettrick water, Afon Ystwyth, River Lune): (a) derived directly in area-by-number (AbN) form and by conversion from area-by-weight (AbW) form using the empirically derived conversion factor of –2.90; (b) derived directly in area-by-weight form and by conversion from area-by-number form using the empirically derived conversion factor of 2.90; (c) in area-by-number form and by conversion from area-by-number form using the theoretically derived conversion factor of –3; (d) derived directly in area-by-weight form and by conversion from area-by-number form using the theoretically derived conversion factor of 3. In all parts, the solid line represents the line of equality and the dashed line indicates the bias. Source: Graham *et al*, 2012. Reproduced with permission of ASCE Library.

the probes into the bed can disrupt the existing stratification (Beschta and Jackson 1979).

Freeze core samples tend to have a 'ragged edge', with larger particles protruding from the frozen mass (Fig. 13.14), implying that all fractions of the distribution are not sampled proportionately. Most importantly, however, freeze core samples are typically less than 10 kg, too small to represent accurately gravels that include particles of size 64 mm and greater (Church *et al.* 1987).

Comparing bulk core and freeze-core sampling

The 'ragged edge' of freeze-core samples would imply that these samples would have fewer fines than bulk core samples of the same gravels. However, comparisons of the two methods by various authors have yielded mixed results, with some studies showing freeze-core samples to be finer, some coarser. In a systematic comparison of shovel, bulk core and freeze-core sampling, Young *et al.* (1991) found that the bulk core samples most frequently approximated the true substrate composition. Bulk core sampling is simple (although labour intensive), can yield large samples and does not suffer from the 'ragged edge' of freeze-core sampling. Therefore, for most purposes, the bulk core sampling approach is more appropriate. Rood and Church (1994) described a hybrid bulk cylindrical–freeze-core apparatus where the samples were unbiased with respect to grain size distributions.

13.8 Sample size requirements

Adequate sample sizes for bulk gravel samples

The question of an adequate size for volumetric samples of coarse sediments is not new. Wentworth (1926, p. 10) recommended that samples should be 'large enough to include

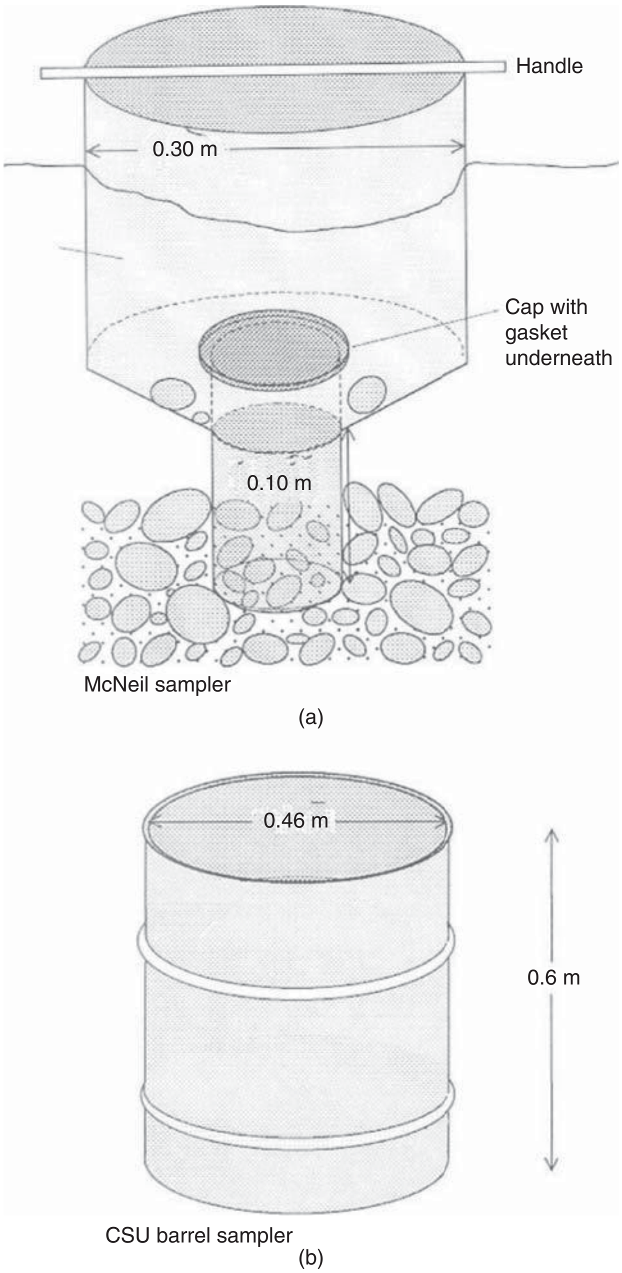


Figure 13.13 (a) Diagram of a simple barrel sampler, made by removing the bottom from a 50 cm (55 gal or 208 L) metal barrel. (b) the FRI (Fisheries Research Institute) sampler has a smaller diameter pipe welded on the bottom and this smaller pipe is driven into the streambed. The sample is removed by hand from the core and the cap on the core retains suspended fine sediment in the basin, to be analysed later for total suspended solids. An alternative method to sample the fine sediment is to collect a depth-integrated sample of the suspended sediment within the basin. (Source: Platts *et al.* (1983).

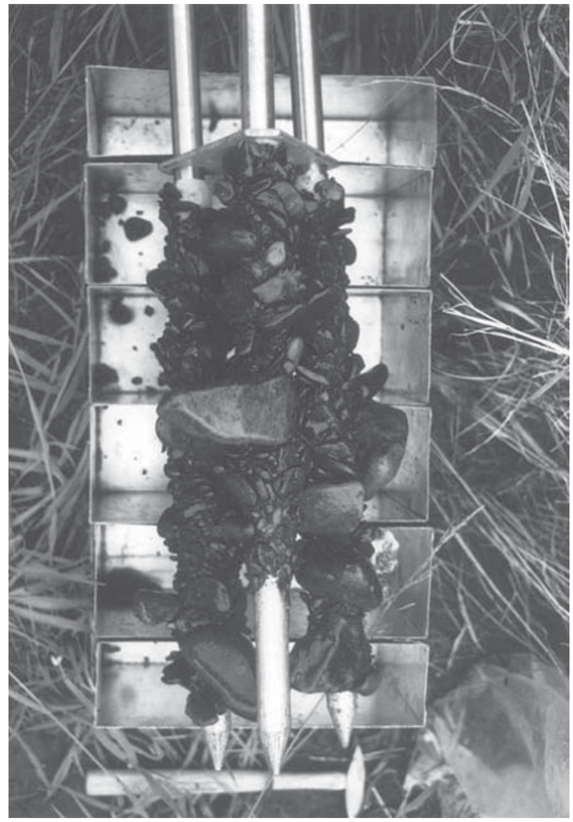


Figure 13.14 Photograph of frozen core extracted from sockeye salmon redd in Quartz Creek, Kenai Peninsula, Alaska. Photograph by G.M. Kondolf, August 1986.

several fragments which fall into the largest grade present in the deposit'. He recognized that '… it is rarely practicable for the geologist to collect samples as large as those demanded by the strict requirements of accuracy' and suggested 'practical' sample sizes along with ideal sample sizes for various grain sizes. For particles 32–64 mm in size, Wentworth's ideal and practical sample sizes were 32 and 16 kg; for particles 64–128 mm in size, the ideal and practical sample sizes were 256 and 32 kg. Mosley and Tinsdale (1985) found the minimum sample size criteria of the American Society for Testing and Materials (ASTM 1978), the British Standards Institution (BSI 1975) and the International Organization for Standardization (ISO 1977) were inconsistent. From 28 bulk samples collected from one location on the Ashley River, New Zealand, they concluded that accurate determination of median size (for this gravel with a D_{50} of 16 mm) required a sample size of 100 kg; provided that the largest stone was less than 5% of the total, the median should be unbiased, but to represent the coarse tail would require larger samples (Mosley and Tinsdale 1985). Church *et al.* (1987) reviewed the sample size problem and concluded that the largest clast should constitute no more than 0.1% of the sample by bulk weight. However, if the largest clast is 100 mm, this would dictate a sample 1300 kg in size. Accordingly, much as Wentworth (1926) presented ideal and practical sizes, Church *et al.* (1987) have, in practice, used the 0.1% criterion for sizes up to 32 mm, thereafter using a 1% criterion up to 128 mm, resulting in samples weighing 150–350 kg.

Sample size has been inadequate in many studies. For example, in spawning gravels of chinook salmon, particles

90 mm or larger are commonly encountered. To represent the coarsest fraction accurately, Wentworth's rule would imply that bulk samples should be about 30 kg, whereas Church *et al.*'s (1987) recommendation would call for samples weighing over 200 kg. Many spawning gravel samples reported in the literature were smaller than 30 kg, and therefore probably do not accurately represent the coarser grades. However, the egg pockets in chinook salmon redds, and the entire redd of smaller species, may consist of considerably less than 30 kg of gravel, so larger samples would, by necessity, include particles from outside the egg pocket or redd. This raises the question of what is being sampled. In the case of a small egg pocket, the entire population may be obtainable, so the sample size criteria, which were designed to obtain representative samples from an unobtainable population, become irrelevant. Platts and Penton (1980) utilized multiple freeze-core probes and heavy equipment to sample an entire, large redd, but this approach is hardly practical for most studies. One approach to this problem is to lump small samples from many redds together into a large, composite sample. This procedure would mask variability in gravel size among redds, but much of the apparent variability may be due to problems in representing the larger size fractions in small samples, so composite size distributions may be more accurate measures of the population. Many studies of spawning gravel have reported only averaged, composite size distributions. If obtained from relatively homogeneous stream channel conditions, these composites probably reflect the gravel population well, but composites from a variety of channel types will not reflect the actual population at any site.

Sample size can affect the size distribution obtained, with larger D_{50}s obtained from larger samples (Ferguson and Paola 1997). In gravels with a few large particles distributed throughout, the size distribution of a given sample may look very different depending on whether it happened to include one of those large rocks. For example, a single 150 mm rock might constitute 20% of the entire sample. In such a sample, percentage values for the other grades would be decreased by one-fifth if the large particle were included or increased by one-quarter if it were excluded. The influence of occasional large particles on the values of other size grades in spawning gravels has been widely recognized and many authors in the fisheries literature have dealt with the problem by excluding large particles from the analysis, such as Chambers *et al.* (1954, 1955), who excluded rocks larger than 152 mm, McNeil and Ahnell (1964), 102 mm, Adams and Beschta (1980), 51 mm, and Tappel and Bjornn (1983), who found that size distribution curves plotted straighter if rocks > 25 mm were excluded. The implications of excluding large rocks depend on what is to be done with the data. If the study is designed to document variations in fine sediment content over space or time, the approach can be justified as an alternative to collecting impractically large samples. However, size data drawn from such truncated curves may not accurately reflect framework sizes used by fish and they certainly will not accurately reflect grain roughness. Similarly, computation of the percentage of fine material will also be affected by truncation. If the implications of truncation are not explicitly recognized, results from one study may be misapplied to another site.

Church *et al.* (1987) recommended that grain size distributions should be computed and compared only for the ranges that have been representatively sampled. For pebble counts, this implies a lower truncation point (e.g. 4 or 8 mm); for bulk samples, it implies an upper truncation point that is a function of sample size and the standard selected.

The large bulk samples needed to satisfy sample size requirements mean that it becomes unwieldy to bring the adequately sized samples (hundreds of kilograms) back to the laboratory to sieve. Therefore, some field sieving is usually necessary. Typically, the procedure is to extract, sun dry and weigh a large sample, sieve and weigh all the fractions coarser than a threshold size such as 8 or 11.2 mm, and split and weigh the (well-mixed) remaining sample until a subsample of a few kilograms remains, which is taken back to the laboratory to run through finer sieves. For large samples, more than one splitting and weighing step can be performed in the field. Initially, all of the largest rocks are individually passed through the template and the remainder of the sample is either split and sieved or all of it is sieved by passing the sample through rocker sieves with large screens (typically with a sieve size up to 64 mm) down to the size threshold. Dry sieving works well in warm, dry weather conditions, where gravels can be sun dried, but may be impractical in wet conditions. Wet sieving is an alternative in wet weather or to process large volumes quickly, although we find it troublesome to handle the fine sediment and water mixture. (Doing field work in Alaska, drying gravels in buckets over a campfire inspired many curious looks and discreet inquiries, most assuming that extraction of gold was involved!) Wilcock *et al.* (1996b) used wet sieving to process numerous Helley–Smith bedload samples on a raft and to process large (~250 kg) bed material samples.

The large effort in obtaining statistically robust subsurface samples motivates schemes to reduce the size and number of bulk samples for measuring an average particle size distribution for a reach of river. Assuming a correlation between surface and subsurface size distributions, the bed can be stratified according to mapped surface patches, subsurface material sampled in selected patches and a weighted-average size distribution computed (Lisle and Madej 1992). If the average subsurface size distribution underlies the surficial patch having the average surface size distribution (Lisle and Madej 1992), an average subsurface sample can be obtained below the surficial patch type having the average surface size for the reach, although it would be prudent to amalgamate subsamples of subsurface material from a number of locations in the average surficial patch.

Sample size and reproducibility of pebble counts

It was the very large bulk sample sizes indicated for coarse gravel that motivated the development of the pebble count (Wolman 1954) as a way of obtaining a sufficiently large sample without having to collect samples that were impossibly large and heavy. Wolman (1954) found that a count of 100 stones produced consistent median grain sizes for multiple counts by one operator and among different operators. Because pebble counts yield large data sets with minimal field work, they have attracted many studies of inter-operator variability, minimum sample size, etc. (e.g. Brush 1961; Hey and Thorne 1983; Mosley and Tinsdale 1985; Fripp and Diplas 1993; Marcus *et al.* 1995; Rice and Church 1996; Wohl *et al.* 1996; Petrie and Diplas 2000). These studies reached various conclusions about sample size, but we can conclude that 100 particles can reliably characterize central tendency, while larger samples can increase confidence in percentile values near extremes of the distribution. It is unclear to what extent some of the published studies restricted pebble counts to 'homogeneous' populations. If more than one geomorphic feature is included in the sample, the potential for error is greater because of the potential to sample different proportions of two different populations. Therefore, minimum sample recommendations should apply to sampling a single population. Lithologies with distinct densities require individually adequate sample sizes. For example, the bed material of the lower Carmel River, California, consists mostly of granitic and metamorphic clasts with a specific gravity of about 2.7, but clasts of Monterey Formation (a Tertiary marine siltstone) with much lower specific gravity also occur, so pebble counts to characterize this bed material were based on at least 100 non-Monterey clasts (Kondolf and Matthews 1986).

13.9 Comparability of pebble counts and bulk samples

If there were no difference between the surface layer and subsurface size distributions, the surface could be considered a random slice through the deposit and the pebble count would yield a random sampling of grains. The pebble count is a random point sampling procedure, so its results are theoretically equivalent to bulk sampling and sieve analysis for sediments with constant density (Kellerhals and Bray 1971). However, the surface layer is typically deficient in fines relative to the subsurface population. Thus, by virtue of the real differences between surface and subsurface layers in gravel bed rivers, the pebble count is usually sampling a different population than bulk sampling. Wolman (1954) noted that pebble counts tend to yield coarser grain size distributions than bulk samples of the same gravel deposit because the former are commonly deficient in fine sediments. This is illustrated by comparing pebble count and bulk samples for recently deposited gravel and sand on the Middle Yuba River, California (Fig. 13.15). The pebble count and bulk sample distributions deviate at the fine tail, reflecting the deficiency in fine sediment at the surface. If particles smaller than 4 mm are excluded from the bulk sample analysis, the resulting curve tracks the pebble count more closely.

Leopold (1970) proposed that results of pebble count analyses be adjusted to compensate for a 'bias … towards larger sizes which, because of their area, are more likely to be picked up'. However, provided that grain volume is proportional to grain weight (true with constant density), there is no bias in a random point count. If larger grains are more likely to be encountered, it is because they occupy a greater part of the cross-sectional area of the slice (the surface) and thus a greater part of the volume of the three-dimensional deposit. Hence there is no theoretical justification for decreasing the actual percentages observed for larger stones provided that the pebble count is conducted correctly, i.e. if sampling is truly random.

13.10 Sampling strategy

A field scientist embarking on sampling a riverbed is commonly faced with a number of general sampling issues, including how to sample distinct facies in the bank and bed, appropriate sample size, what the sample is expected to represent and tradeoffs between consistency and relevance. These issues present themselves in every field problem and should be addressed separately, rather than simply accepting a pre-packaged sampling protocol. Choosing a method described in a manual and used by predecessors provides some assurance that it has been proven and the data so gathered will be accepted by others. However, the heterogeneity of gravel beds and the variety of problems requiring bed material data have motivated a variety of sampling methods, some of which are still evolving. Sampling gravel beds is notoriously demanding for fluvial geomorphologists, requiring, in some cases, the collection of tons of bed material from various locations in the channel and passing it through numerous sieves of low capacity. Sampling locations must be carefully chosen to target areas relevant to the problem, avoid bias and obtain a large enough sample to reduce the variance of a highly variable population adequately, while dealing with practical problems such as collecting samples under flowing water. These constraints mean that, although a 'quick and dirty' effort is likely to be wasted effort, there is no room for extra work that does not satisfy the purposes of the sampling programme.

A bed sampling programme should pre-eminently meet the needs of the problem for which it is intended. Problems can be of regional or even national scale, but most often they entail a reach, basin or biogeographic region. Of course, internal consistency is necessary for comparing spatial and temporal variations within the scope of a study. However, if agencies press for agency-wide consistency for purposes beyond those of individual studies or programmes, field scientists still have an obligation to understand and select the methods best suited to address the question posed, rather than simply defer to agency guidelines. If the methods are chosen correctly to suit the needs of a particular problem (e.g. fine sediment in spawning gravels, hydraulic roughness or channel mobility), then over time these data will also prove useful for more comprehensive analyses addressing the same sort

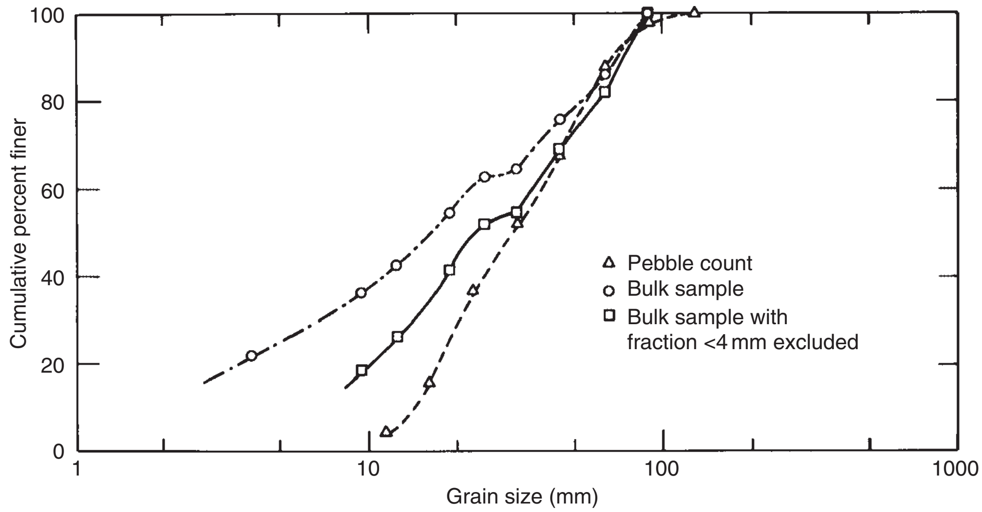


Figure 13.15 Grain size distributions of recently deposited sediment along the Middle Yuba River, California, by pebble count and bulk sample (obtained with shovel from exposed bars). Also shown is size distribution for bulk sample with fraction < 4 mm. Source: Kondolf (1988).

of problem and useful degrees of consistency will naturally arise even as improved methods develop.

The challenge is to array these methods in the field in a sampling scheme that accomplishes the goals of the study. Three qualities should guide the choice of sampling scheme: accuracy, precision and consistency. The obvious reason for accuracy is to measure the part of the river that is meaningful to the problem. To typify spawning habitat, for example, we sample bed material where fish spawn, such as riffle crests. At the same time, we want to avoid bias, perhaps by sampling near actual redds instead of where we predict fish are likely to spawn. Most often, we sample at a number of sites in order to arrive at a representative average of the population to be sampled. These considerations lead to the first essential step – to define the sample population, that is, all of the areas of the riverbed that we want to represent.

The need for precision can also influence the definition of sample population. At the outset, acceptable error should be determined quantitatively or qualitatively from the goals of the study. Sample error is a function of population variance and sample size (Benjamin and Cornell 1970). Population variance can only be limited by redefining a more homogeneous sample population. For example, if one wants to measure downstream fining of sediment inputs or otherwise compare changes in particle size from one reach to another or one river to another, one can sample some consistent hydraulic or sedimentary environments, such as riffle crests or bar heads, and thereby remove some of the variation created by local channel form. Beyond that, sample variance can be reduced only by increases in sample size. In many cases, stratifying the bed according to bed material size and sampling according to these designations can reduce sample error while limiting the number of samples.

Consistency means not only using the same criteria for sample location and measurement within a study, such as exemplified above for downstream fining, but also to match the accuracy, precision and scale of bed material samples to those of other related data. If boundary shear stress is to be measured from reach averages of hydraulic radius and channel gradient, then to compute Shields stress, a composite D_{50} value for the entire bed surface is needed. If the bed is relatively uniform, a single pebble count over the entire bed will be sufficient or, where more than one facies occurs, a weighted average can be computed from grain size distributions measured for individual facies. The error in any one of the three parameters should not greatly outweigh the error of the others.

In practice, these considerations often lead to tensions between objectivity, practicality and professional judgement in choosing sampling schemes and locations. At the risk of bias, an experienced scientist can accurately stratify a bed according to the scale and characteristics of variations that bear on the problem and thereby improve both accuracy and precision of measurements. For example, if one wants an average grain roughness for a reach, then evenly spaced transects over the bed might be appropriate if there is no recognizable organization of bed material into riffles and pools. If some organization is apparent, then placing transects over these distinct features and weighing the data in proportion to the relative area of recognizable strata can lower the variance.

River data, especially bed material size, are messy, and sending legions of novices into the field with written or verbal instructions is doomed to amplify the noise of uncertainty and variability. For this reason, it is essential that lead scientists devote time in the field to choosing the locations of bed material samples, and also the methods, and evaluate if the needs of the study are being met as the data are obtained.

Critique of some popular bed sampling methods

To illustrate these considerations, we evaluate some commonly used bed material sampling schemes according to their application, adequacy of sample size and bias. The 'zig-zag' method (Bevenger and King 1995) called for the observer to walk diagonally along the stream bed, travelling downstream in straight lines from left bank to right bank and back again, 'randomly' selecting 100 pebbles for measurement at a spacing of 3.5 ft (1.1 m). The stated aim of the method was to randomly sample '… numerous meander bends and all associated habitat features … as an integrated unit rather than as individual cross sections', implying that a composite grain size for the entire reach was sought. The zig-zag method lumps data points from the bank and bed features, each of which would have distinct grain size populations. Thus, to a geomorphologist, this method has the fatal flaw of 'mixing apples and oranges'.

The zig-zag method was proposed as a tool to evaluate fine sediment content (and its changes over time) in relation to upstream land use (Bevinger and King 1995), but it has several drawbacks as a monitoring tool. First, the sample size is inadequate to yield accurate size data and develop complete size distribution curves. (The sample size of 100 pebbles specified by Wolman 1954 was for a homogeneous area of bed material, not a mish-mash of bed and bank materials.) Not surprisingly, field tests showed that the zig-zag counts were not reproducible by different observers (Bevenger and King 1995). If it were used to assess fine sediment variations over time, the method would suffer from three sources of variation that could not be resolved: (i) the actual particle size of bed and bank material could vary between surveys; (ii) the relative areas of beds and banks could vary due to bank erosion or accretion; (iii) the relative sampled areas of bed and banks could vary between measurements and it is unlikely that these areas will be sampled according to their areal extent by accident of randomly selecting zigs and zags. The inability to decipher the added noise due to multiple sources of variation would severely hamper the use of this method as a monitoring tool.

Similar issues arise with bank-to-bank transects specified to measure bed material size as input to the Rosgen (1996) classification system. Along a reach 20 channel widths in length, 10 cross-sections are located from bank to bank, along each of which 10 points are randomly sampled for a total sample size of 100. Cross-sections are distributed among riffles and pools in proportion to the abundance of riffles and pools in the reach, so if riffles make up 20% of the reach length, two of the 10 cross-sections are located in riffles. The samples begin and end with the 'bankfull' bank top along each cross-section. This method is an improvement over the zig-zag method in that that a number of transects are placed over pools and riffles according to their relative area in the channel. However, while pools will tend to have distinctly smaller grain sizes than riffles, the differences between pool and riffle grain sizes will not be consistent, and a more accurate stratification could be done by directly mapping surface facies. More serious drawbacks are the small sample size (100 counts in total over the mix of different geomorphic features) and combining bed and bank materials in a count, similar to the zig-zag count. The resulting size distribution would probably have limited application other than as an input to the stream classification.

13.11 Applications of bed sediment sampling related to aquatic habitat

Measurement of fine sediment accumulation in pools: V^*

Fine-grained bed material can be winnowed from the bed surface and accumulate during low flow in pools, where it can form thick patches, reduce pool habitat and affect benthic and intergravel habitats. Residual pool volume is the volume of a pool (disregarding fine bed material) below the elevation of the downstream riffle crest (Bathurst 1981). The fraction of residual pool volume filled with fine bed material (V^*) was developed as a measure of the in-channel supply of excess fine bed material (Lisle and Hilton 1992, 1999). The particle size of fine sediment in pools varies among channels, but in our experience in many streams the size usually ranges from fine sand to fine gravel.

V^* is a dimensionless parameter that is essentially independent of pool size, best measured during low flow when the water surface of the pool is nearly horizontal (an assumption in the calculations), the channel can be easily waded or navigated and the bed is visible. In each pool, water depths are sounded and the thickness of fine material is probed with a graduated steel rod at roughly 50 locations total along 4–8 transects across the pool, depending on the complexity of pool topography and the distribution of fine patches (Fig. 13.16) (Hilton and Lisle 1993). Volumes of water and fine material are computed within the boundaries of the residual pool by summing the volumes contained between adjacent transects, calculating V^* on-site with a program available at www.fs.fed.us/psw/topics/water/vstar/. An experienced team of three can measure and compute a value of V^* in a pool in a wadeable channel in approximately 1 hour.

Like many parameters for natural channels, V^* can be highly variable, so 8–20 pools (depending on the variation in V^* in the reach) must be measured to obtain a reach-mean value of V^*. The most effective and statistically powerful application of V^* is for monitoring in a reach of channel, where factors other than sediment supply (e.g. flow regime, lithology) remain essentially constant. Fine bed material stored on the bed surface has very short residence times and changes in supply from the basin register quickly as changes in storage in pools (Lisle and Hilton 1999). High values of V^* signify large chronic or recent inputs of fine sediment to a channel.

Because a number of factors affect V^*, when using V^* to interpret channel condition supporting information should be obtained about basin conditions, such as basin lithology (V^* is most applicable in basins whose lithology produces abundant fine sediment); flow regime (snowmelt-dominated regimes can be expected to flush fine sediment from the bed surface and have low values of V^*) and sediment sorting (the method is applicable to armoured channels where the interface between a layer of fine sediment overlying a gravel bed can be detected with confidence) (Lisle and Hilton 1999).

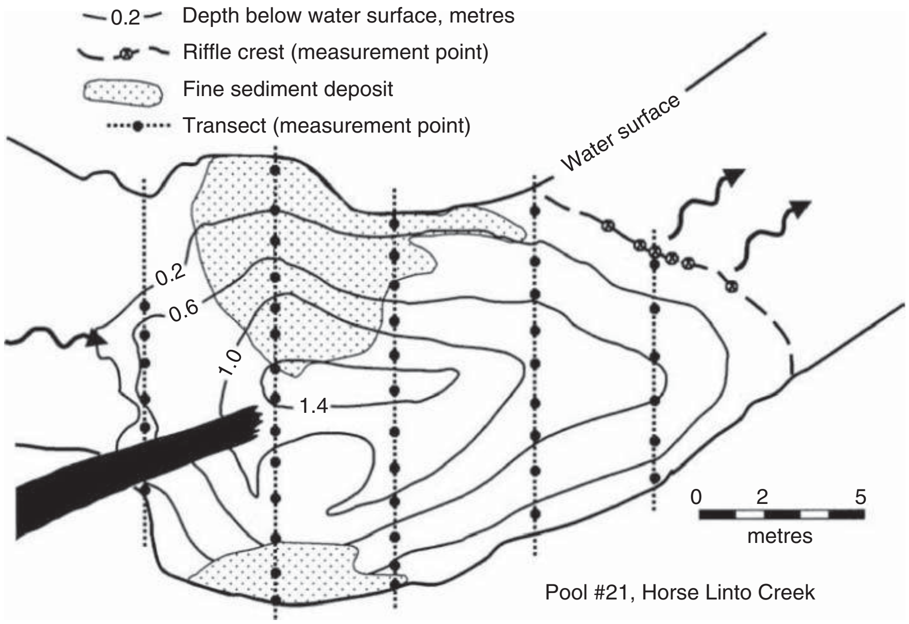


Figure 13.16 Measuring fine sediment in pools to calculate V^*. Water depth is measured along transects in the pool and subtracted from water depths at the riffle crest to compute residual depths. Fine sediment thicknesses are probed along the same transects and, in some cases, augmented by more points over the deposits. Residual pool volume and fine-sediment volume are then computed. Source: Hilton and Lisle (1993); www.fs.fed.us/psw/topics/water/vstar/.

Assessing salmonid spawning gravel quality

The size of available streambed gravels can limit the success of spawning by salmonids (Groot and Margolis 1991). The bed material may be too coarse for spawning fish to move, a common problem below dams, or excessive interstitial fine sediment may clog spawning gravels, as documented downstream of land uses that increase sediment yields, such as timber harvest and road construction (Cederholm and Salo 1979; Everest *et al.* 1987; Meehan 1991).

Because of these problems, there is frequently a need to assess the quality of spawning gravels to determine whether gravel size limits spawning success. Any such assessment involves comparison of gravel size on-site with information on gravel size suitability from laboratory studies or field observations elsewhere. Despite the literature devoted to the search for a single statistic drawn or computed from the streambed particle size distribution to serve as an index of gravel quality (e.g. Lotspeich and Everest 1981; Shirazi and Seim 1981, 1982; Beschta 1982), a natural gravel mixture cannot be fully described by any single statistic, and because gravel requirements of salmonids differ with life stage, the appropriate descriptor will vary with the functions of gravel at each life stage (Kondolf 2000).

To assess whether gravels are small enough to be moved by a given salmonid to construct a redd, the size of the framework gravels is of interest and the D_{50} or D_{84} should be compared with the spawning gravel sizes observed for the species elsewhere. To assess whether the interstitial fine sediment content is so high as to interfere with incubation or emergence, the percentage of fine sediment of the potential spawning gravel should be adjusted for probable cleansing effects during redd construction and then compared with rough standards drawn from laboratory and field studies of incubation and emergence success. An assessment should also consider that the fine sediment content of gravel can increase during incubation by infiltration, the gravels may become armoured over time or downwelling and upwelling currents may be inadequate. These considerations are incorporated in a nine-step, life-stage-specific assessment approach (Fig. 13.17), described in detail in Kondolf (2000).

Measuring infiltration of fine sediment into spawning gravels

Whatever the concentration of fine sediment, the available evidence suggests that fish can typically flush enough from spawning gravel during redd construction to provide initially adequate intergravel flow of oxygenated water to incubating embryos (Kondolf *et al.* 1993). Fine sediment transported by subsequent flows, however, can infiltrate and fill intergravel pores (e.g. Carling and McCahon 1987) and higher flows that mobilize the gravel can deposit new layers of bed material containing abundant fines (Lisle 1989). Therefore, the critical measure of spawning habitat is not the initial size composition of spawning gravel, but the gravel composition during the incubation period. To measure changes in gravel composition caused by sediment-transporting events, containers (buckets or cans) can be filled with clean gravel and buried flush with the bed surface (Lisle and Eads 1991). The containers are retrieved after the allotted time and their contents sieved in order to measure the volume of infiltrated sediment. This method is most appropriate where scour and fill of the bed are minimal and most infiltrating sediment (fine sand or larger) is coarse

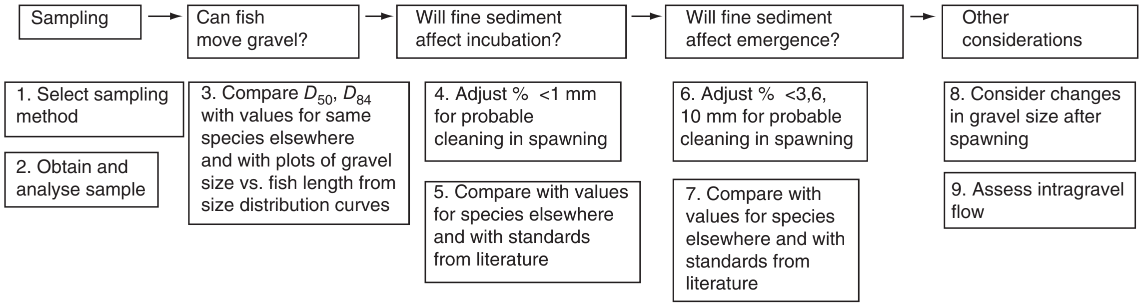


Figure 13.17 Flow chart illustrating nine discrete steps in evaluating salmonid spawning gravel quality. Source: Kondolf, 2000. Reproduced with permission of Taylor and Francis.

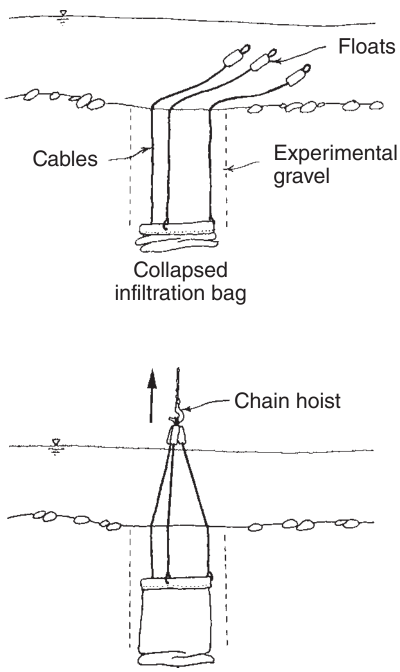


Figure 13.18 The collapsed-bag technique for obtaining an unbounded sampled of bed material infiltrated by fine sediment. Source: Lisle and Eads (1991).

enough to be predominantly influenced by gravity rather than intergravel flow once it penetrates the bed surface. The containers can be solid walled or porous, with the difference being that solid-walled containers do not receive fine-grained sediment carried by intergravel flow, but also do not lose any infiltrated sediment to intergravel flow.

Another method solves the problems of intergravel transport and scour and fill, but installation is more demanding (Lisle and Eads,1991; George Sterling, University of Alberta, personal communication). A collapsed bag sewn onto a steel rim is buried under an unbounded column of clean gravel and later pulled vertically out of the bed, enclosing the overlying infiltrated gravel (Fig. 13.18). The bed is first excavated from inside an open cylinder down to the desired depth. The collapsed bag is placed open end up in the pit and cables attached to the rim are extended to the surface. Clean gravel is poured back into the pit, the cylinder removed and the surface layer replaced. To retrieve the sample, the cables are drawn upwards with a chain hoist mounted overhead. In the process, the bag rises rim first through the gravel, capturing the sample. Scour chains can be installed alongside to measure the contribution of scour and fill to changes in gravel composition and freeze tubes can be installed within the sample to measure the stratigraphy of fine deposits. As in the other methods, the sample is sieved to measure the influx of fine sediment.

13.12 Case study: determining changes in fine sediment content during flushing flows, Trinity River, California

Description of case study site

Since construction of the Trinity and Lewiston Dams in 1961, about 80% of runoff from the upper Trinity River has been exported to the Sacramento River, reducing high flows such that fine sediment delivered from tributaries accumulated in the bed of the Trinity without being flushed out, degrading spawning gravels and other habitats for anadromous fish. As part of a legally-mandated effort to restore fish populations in the Trinity River, the US Bureau of Reclamation made a series of controlled, experimental, high-flow releases ('flushing flows') from Trinity and Lewiston Dams in each of three years, 1991, 1992 and 1993: 76, 164 and 80 $m^3\ s^{-1}$ (USFWS 1999). The 80 and 164 $m^3\ s^{-1}$ were both well below the $Q_{1.5}$ on the pre-dam flood frequency curve, but correspond to approximately the Q_3 and Q_7 on the post-dam curve (USFWS 1999). Wilcock *et al.* (1995, 1996a,1996b) documented the effect of the flushing flows on spawning gravel quality and channel form at two study reaches heavily used by spawning salmon, Poker Bar and Steelbridge (Fig. 13.19).

Case study methods and results

During the flushing flows, Wilcock *et al.* (1996a,1996b) measured vertical velocity profiles and sampled bedload in transport. Before and after the flushing flows they placed tracer gravels and bedload traps to document bed mobility, surveyed

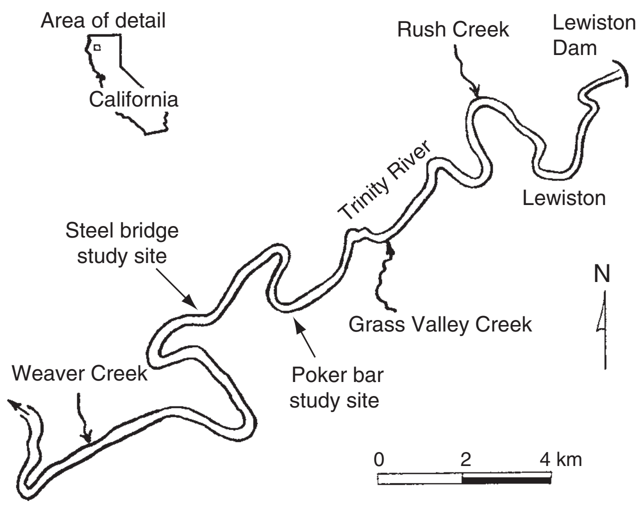


Figure 13.19 Location map of the Trinity River downstream of Trinity and Lewiston Dams, showing study reach of Wilcock *et al.* (1995, 1996a,1996b). Source: Wilcock *et al*, 1996. Reproduced with permission of AGU.

cross-sections, visually estimated fine sediment stored in the bed surface over a 3 km study reach and measured changes in fine sediment in the bed at the study sites through repeated visual observation, pebble counts and bulk core sampling. They made measurements during the 1993 flow only at the Poker Bar site, as this flow was essentially the same as the 1991 flow.

Visual estimates of surficial fine sediment in a 3 km reach

Wilcock *et al.* floated the reach from the confluence of Grass Valley Creek to the Steelbridge study site, visually estimating the percentage of fine sediment on the bed, before the 1992 release and after the 1993 release. They chose this approach as the only practicable way to quantify (albeit roughly) the volume of fine sediment stored in the bed elsewhere in the study reach, information needed for overall sediment routing calculations. They computed sediment storage values for each of six subreaches, bounded by large pools (some of which have been dredged to reduce the river's sand load). After the 1993 release, fine sediment percentages were reduced in all subreaches.

Visual estimates and pebble counts at detailed study sites

In 1991, Wilcock *et al.* visually estimated variations in bed roughness for hydraulic modelling by classifying the sediment as sand, gravel, cobble or boulder. In 1992; they expanded the visual estimates with the goal of detecting changes in bed texture (especially fine sediment content) and (at regular transect points) estimated the percentage of the bed covered by sediment < 8 mm (commonly termed 'percent embedded' in the fisheries literature), and also the D_{50} and D_{90} (to the nearest Wentworth size class) and used the same operator for all observations. The estimated error was ±10% for percentage embedded and ±one size interval (i.e. *phi*) unit for D_{50} and D_{90}. To characterize surficial sediment, Wilcock *et al.* conducted pebble counts along the cross-sections, measuring 100 stones per cross-section in 1991 and 200 stones per cross-section in 1992 and 1993. The bed material was relatively consistent across the channel in these study reaches, justifying the use of a single pebble count for a section-wide characterization. The 1991 release was too small to produce a significant change in the substrate, but the 1992 release decreased surficial fine sediment at nearly all cross-sections, as measured by visual estimates of percentage < 8 mm and by pebble count (Fig. 13.20).

Bulk sampling at detailed study sites

To quantify better the changes in fine sediment content at the study sites, Wilcock *et al.* collected three types of bulk samples from the cross-sections used annually by spawning salmon: pre-release, post-release at pre-release locations and new post-release samples next to the original samples, the last to control for the effect of pre-release sampling on the post-release sediment composition. They also inserted tracer gravels in the pre-release sample sites (see Chapter 10 for a review of tracer gravel tools). After surveying the bed elevation at the sample point, they inserted a metal cylinder into the bed as deep as possible and removed all sediment down to the bottom of the sampler, using (in 1992) a 59 cm diameter cylinder (a bottomless 55 gallon drum) and sampling as deep as 40 cm, to yield sample sizes of 112–281 kg (mean 182 kg). To process these large samples, they wet sieved on-site, counting coarse particles (> 8 mm) and measuring the volume of sediment < 8 mm, converting to mass based on relations established the previous year.

Bulk sample results were less consistent than the surface sampling results. For example, at Poker Bar cross-section 2, the pre-release percentage < 8 mm ranged from 23 to 34% and the post-release from 26 to 35%. Results from various methods of sediment sampling are summarized in Fig. 13.21, displaying the range of results possible from these diverse methods. The Trinity River study illustrates how a wide range of bed material sampling approaches can be combined to provide a multifaceted picture of bed sediment change, in this case resulting from experimental flow releases (Wilcock *et al.* 1996a,1996b).

13.13 Case study: application of *V** to French and Bear Creeks, California

*V** is most applicable to monitoring annual variations in a reach of channel and for evaluating the transport and spread of well-defined inputs of fine sediment. Two examples from northwestern California illustrate these uses:

French Creek

French Creek drains 60.4 km^2 in the Klamath Mountains, flowing into the Scott River near Etna, California. Much of

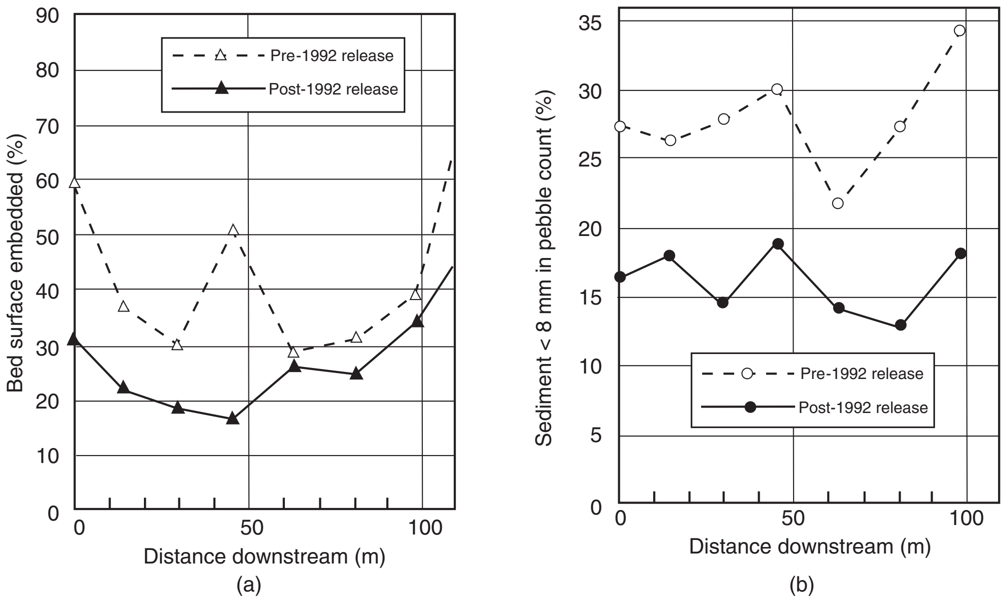


Figure 13.20 Fine sediment in the bed versus downstream distance at Poker Bar study site on the Trinity River in 1992, before and after experimental flushing flows. The open symbols and dashed lines represent pre-release conditions and the closed symbols and solid lines represent the post-release conditions. (a) Visually estimated embeddedness or percentage of surface covered by fine sediment. (b) Percentage finer than 8 mm as measured in pebble counts. Source: Wilcock *et al*, 1995. Reproduced with permission of the Center for Environmental Design Research, University of California at Berkeley.

its basin is underlain by deeply weathered granitic soils that erode to sand-sized sediments. Erosion of logged areas before 1990 contributed large volumes of fine sediment to the channel, filling pools and armour interstices in riffles with coarse sand and negatively affecting habitat for a native population of anadromous salmonids. From 1991 to 1994, landowners collaborated with the Klamath National Forest on an erosion control programme to reduce sediment supply. Because the problem was sand entering a gravel bed channel, V^* was selected to test the effectiveness of the programme to reduce the in-channel supply of sand. The monitoring programme eventually focused on a mainstem reach including 10 pools not far downstream of the sediment sources.

Soon after the erosion control programme was implemented in 1992, the fines volume decreased by more than half as scoured-pool volume (residual volume minus fine sediment volume) remained essentially unchanged (Fig. 13.22). Values of V^* decreased to approximately one-third the initial value. A large rain-generated flood in January 1997 delivered fine sediment to the channel and caused fines volume and V^* to nearly double, but in subsequent years V^* decreased again and pool volume recovered. The background value of V^* in French Creek as of 2001 (≤ 0.1) was equal to the reference value for channels draining weathered granite (Lisle and Hilton 1999). Thus, the V^* values provided strong evidence for the effectiveness of the erosion control programme: in-channel supplies of sand (the dominant sediment input) decreased to background levels.

Bear Creek

Bear Creek drains a 20 km^2 basin, also in the Klamath Mountains, and enters the South Fork Trinity River near Hyampom, California. Like French Creek, it is underlain predominantly by weathered granite, but unlike French Creek, its basin is nearly undisturbed. In 1991; V^* was measured in a reach of channel including 19 pools to help establish reference values for V^*. In the upstream portion of the reach, V^* was indeed low (< 0.1), but at the half-way point, V^* increased dramatically to > 0.5 and then decreased downstream (Fig. 13.23). Upon further inspection, Lisle and Hilton (1992) discovered the input of fine sediment from a small, illegal mining operation upslope from the point of the large inflection of V^*. Although it was not the purpose of the study, the V^* values permitted the detection of a sediment source and evaluation of its effect on the channel.

13.14 Conclusion: selecting an appropriate sampling method

Let the punishment fit the crime! This chapter's overriding message is that the purposes of the study must be clearly articulated and the methods chosen should logically follow from the questions posed. When selecting a sampling method, we should ask ourselves, why are we collecting these data? Precisely how do we plan to use them? And what confidence limits are required to answer the questions we pose of the data?

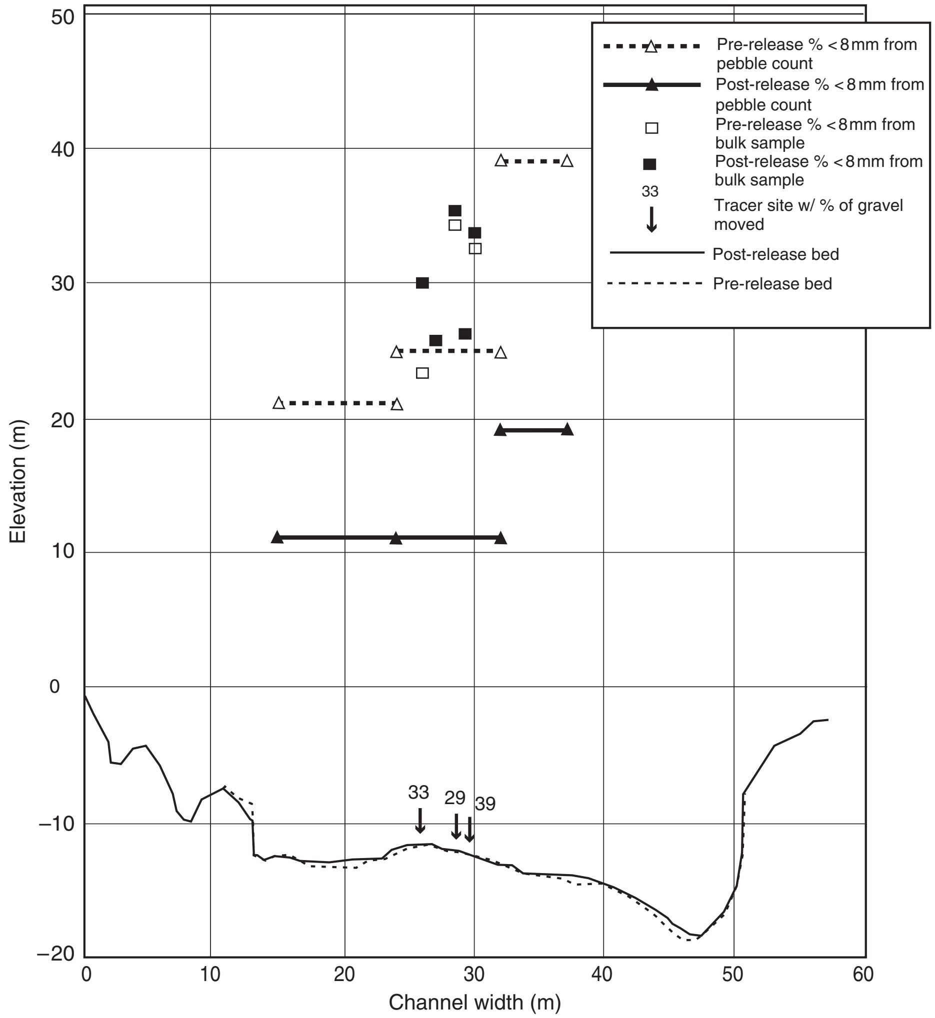


Figure 13.21 Bed elevation and sediment conditions at cross-section 2 of the Poker Bar study site on the Trinity River, before and after experimental flushing flow releases in 1992. Dashed lines and open symbols denote pre-release conditions; solid lines and filled symbols denote post-release conditions. Pre- and post-release cross-sections are shown in metres relative to an arbitrary datum and have been vertically exaggerated five times. Horizontal lines represent pebble counts, with the extent of the horizontal line indicating the cross-sectional extent of facies sampled and vertical position reflecting the percentage less than 8 mm. Percentages less than 8 mm are also shown for all bulk samples: four pre-release samples and six post-release samples. The location of the tracer sites are shown with percentage tracer (by mass) which moved during the release. Source: Wilcock *et al*, 1995. Reproduced with permission of the Center for Environmental Design Research, University of California at Berkeley.

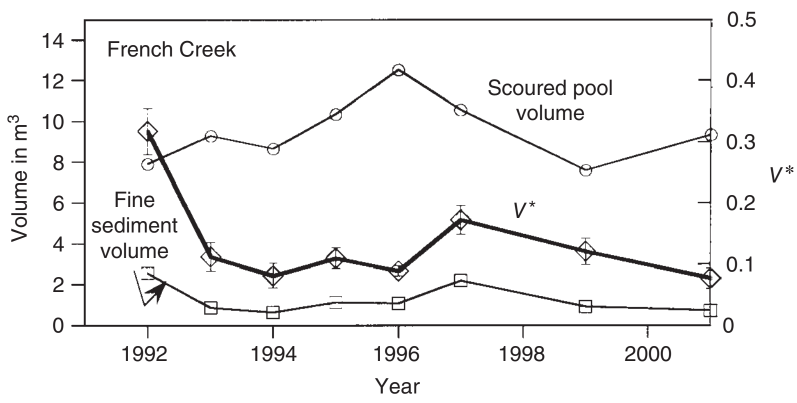


Figure 13.22 Annual variations in V^*, fine-sediment volume and scoured-pool volume in French Creek, 1991–2001. Source: Lisle and Hilton, 1999. Reproduced with permission of Elsevier.

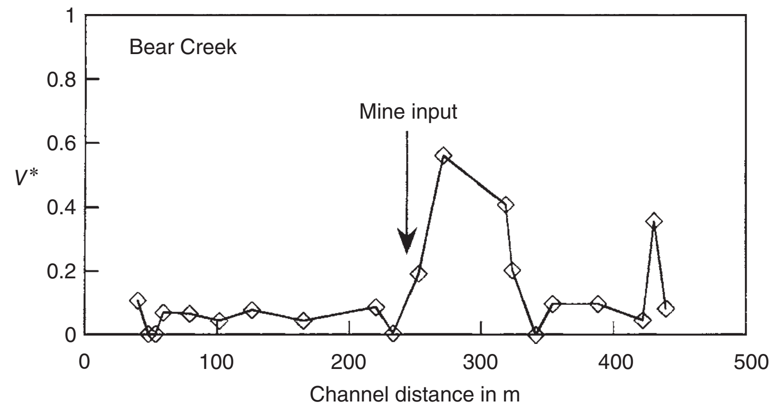


Figure 13.23 Variations in V^* with longitudinal distance along Bear Creek, showing the influence of previously unknown, illegal mining operation on fine-sediment loading. Source: Lisle and Hilton, 1992. Reproduced with permission of Elsevier.

Table 13.2 Advantages and disadvantages of various tools for sampling bed material.

Objective Estimate grain roughness	**Method** Pebble count or image-based methods	**Advantages/disadvantages** Produces reproducible estimate of surface grain size	**Reference** Wolman 1954; Carbonneau *et al.* 2005; Lejot *et al.* 2011
Calculate bed mobility threshold	Pebble count or image-based methods	The grain size of the surface layer controlling bed mobilization can be assessed via pebble count or image-based methods	Kondolf 1997
Calculate bedload transport assuming subsurface sediment reflects bedload	Bulk sampling	If subsurface sediment reflect bedload in transit, subsurface sample needed. Captures more of the mobile fine-grained tail where present	Parker and Klingeman 1982
Calculate bedload transport without assuming subsurface reflects bedload	Pebble count or image-based methods	Where the relation between subsurface grain sizes distribution and size of bedload in transit is unknown, can use surface-based equation. This may underestimate transport by missing finer fraction where present	Parker 1990
Assess suitability of framework size of salmonid spawning gravels	Pebble count or image-based methods, or bulk sample	Pebble count and image-based methods are faster than subsurface samples and provide reasonable estimates.	Kondolf 2000
Assess interstitial fine sediment content of spawning gravels	Bulk sample	Provides information on fine sediment content not visible from surface inspection alone (gravels that appear 'clean' on the surface may contain high interstitial fine sediment subsurface)	Kondolf 2000
Assess vertical distribution of fine sediments in gravel	Freeze-core sample	Provides vertical stratification, but sample size is small and stratification may be disrupted by insertion of probe	Adams and Beschta 1980; Everest *et al.* 1980

Adapted from Wentworth, 1926.

As suggested in Table 13.2, different sampling methods lend themselves to different questions. For example, the pebble count (Wolman 1954) was developed to quantify grain size for surface roughness, to avoid the need for inconveniently large bulk samples. It is a simple and useful technique for measuring the size distribution of the surface layer (e.g. for estimating grain roughness, predicting bed mobilization thresholds, assessing framework size of spawning gravels or tracking changes in surficial fine sediment content in specific geomorphic channel units), but it does not address subsurface size distributions (including interstitial fine sediment content), nor does it address the distribution of surficial sediments beyond the unit sampled (e.g. the bar, riffle or other geomorphic feature on which the pebble count is performed).

To calculate bedload transport, subsurface bulk samples are needed if we assume that the subsurface grain size distribution approximates that of the bedload. However, obtaining these is labour intensive and cannot be done over the entire bed. Moreover, since there is uncertainty whether the subsurface material in a particular channel does in fact represent bed load or if it is too coarse, an alternative is to sample the surface layer and use Parker's (1990) surface-based equation.

To assess the interstitial fine sediment content of gravels to assess the quality of spawning gravels (or for other ecological studies) requires subsurface bulk samples. Again, there is the problem that obtaining these is too labour intensive to do everywhere, so it might be done in specific units such as important spawning gravels and assumed to apply to other spawning beds. However, in other cases there is need to extrapolate these results over a broader reach of river. Rather than assume that the unsampled 99.9% of the channel bed is represented by the 0.1% sampled, a very different approach is to map the facies (as in Fig. 13.7), one providing less detailed information (no size distribution curves) about any one point, but useful information over a wide area. In the Oregon Branch example in Wolman and Schick (1967), the question concerned the effects of increased sediment yield from part of the catchment and the facies map showed where in the channel the sediment deposited and the proportion of the bed surface covered by fine sediment. It can be particularly powerful to combine facies mapping with pebble counts or bulk sampling of specific facies units, with the former providing an overall context and the latter providing specific size distribution data.

Acknowledgement

We are grateful for the contributions of the late M. Gordon Wolman to this chapter as it appeared in first edition of the book, especially his wonderful summing up of the problem, *Let the punishment fit the crime!*

References

Adams, J. 1979. Gravel size analysis from photographs. *Journal of the Hydraulics Division, American Society of Civil Engineers* 105: 1247–1285.

Adams, J.N. and Beschta, R.L. 1980. Gravel bed composition in Oregon coastal streams. *Canadian Journal of Fisheries and Aquatic Sciences* 37: 1514–1521.

ASTM. 1978. *Standard Methods of Sampling Aggregates, ANSI/ASTM D75-71*, Philadelphia, PA: American Society of Testing and Materials.

Bathurst, J.C. 1981. Discussion of bar resistance of gravel-bed streams. *Journal of the Hydraulics Division, American Society of Civil Engineers* 107(HY10), 1276–1278.

Benjamin, J.R. and Cornell, C.A. 1970. *Probability, Statistics and Decision for Civil Engineers*, New York: McGraw-Hill, 682pp.

Bertoldi, W., Piégay, H., Buffin-Bélanger, T., Graham, D. and Rice, S. 2012. Applications of close-range imagery in river research. In: Carbonneau, P.E. and Piégay, H., eds., *Fluvial Remote Sensing for Science and Management*, Chichester: John Wiley & Sons, pp. 341–366.

Beschta, R.L. 1982. Comment on 'stream system evaluation with emphasis on spawning habitat for salmonids' by Mostafa A. Shirazi and Wayne K. Seim. *Water Resources Research* 18: 1292–1295.

Beschta, R.L. and Jackson, W.L. 1979. The intrusion of fine sediments into a stable gravel bed. *Journal of the Fisheries Research Board of Canada* 36: 204–210.

Bevenger, G.S. and King, R.M. 1995. *A Pebble Count Procedure for Assessing Watershed Cumulative Effects, Research Paper RM-RP-319*, Fort Collins, CO: US Department of Agriculture, Forest Service, Rocky Mountain Forest and Range Experiment Station.

Black, M., Carbonneau, P., Church, M. and Warburton, J. 2014. Mapping sub-pixel fluvial grain sizes with hyperspatial imagery. *Sedimentology* 61: 691–711.

Bovee, K.D. 1982. *A Guide to Stream Habitat Analysis Using the Instream Flow Incremental Methodology, Instream Flow Information Paper 12, FWS/OBS-82/86*, Washington, DC: US Fish and Wildlife Service.

Brush, L.M. 1961. *Drainage Basins, Channels and Flow Characteristics of Selected Streams in Central Pennsylvania, US Geological Survey Professional Paper 282-F*, Washington, DC: US Government Printing Office.

BSI. 1975. *British Standard Methods for Sampling and Testing of Mineral Aggregates, Sands and Fillers. Part I. Sampling, Size, Shape and Classification, BS 812 Part I: 1975*, London: British Standards Institution.

Buffington, J.M. and Montgomery, D.R. 1999a. A procedure for classifying textural facies in gravel-bed rivers. *Water Resources Research* 35(6): 1903–1914.

Buffington, J.M. and Montgomery, D.R. 1999b Effects of sediment supply on surface textures of gravel-bed rivers. *Water Resources Research* 35(11): 3523–3530.

Bunte, K. and Abt, S.R. 2001. *Sampling Surface and Subsurface Particle-Size Distributions in Wadable Gravel- and Cobble-Bed Streams for Analyses in Sediment Transport, Hydraulics, and Streambed Monitoring, General Technical Report RMRSS-GTR-74*, Fort Collins, CO: US Department of Agriculture, Forest Service, Rocky Mountain Research Station.

Burns, D.C. 1978. Spawning bed sedimentation studies in northern California streams. *California Fish and Game* 56: 253–270.

Buscombe, D., Rubin, D.M. and Warrick, J.A. 2010. A universal approximation of grain size from images of non cohesive sediment. *Journal of Geophysical Research* 115: F02015.

Carbonneau, P.E., Bergeron, N.E. and Lane, S.N. 2005. Automated grain size measurements from airborne remote sensing for long profile measurements of fluvial grain sizes, *Water Resources Research* 41: W11426.

Carling, P.A. and McCahon, C.P. 1987. Natural siltations of brown trout (Salmo trutta) spawning gravels during low-flow conditions. In: Craig, J.F. and Kemper, J.B., eds., *Regulated Streams: Advances in Ecology*, New York: Plenum Press, pp. 229–244.

Carling, P.A. and Reader, N.A. 1982. Structure, composition and bulk properties of upland stream gravels. *Earth Surface Processes and Landforms* 7: 349–365.

Cederholm, C.J. and Salo, E.O. 1979. *The Effects of Logging Road Landslide Siltation on the Salmon and Trout Spawning Gravels of Stequaleho Creek and the Clearwater River Basin, Jefferson County, Washington, 1972–1978, Report No. FRI-UW-7915*, Seattle, WA: Fisheries Research Institute, University of Washington.

Chambers, J.S., Pressey, R.T., Donaldson, J.R. and McKinley, W.R. 1954. *Research Relating to Study of Spawning Grounds in Natural Areas, Annual Report to US Army Corps of Engineers, Contract DA-35026-Eng-20572*, Olympia, WA: State of Washington, Department of Fisheries.

Chambers, J.S., Allen, G.H. and Pressey, R.T. 1955. *Research Relating to Study of Spawning Grounds in Natural Areas, Annual Report to US Army Corps of Engineers, Contract DA-35026-Eng-20572*, Olympia, WA: State of Washington, Department of Fisheries.

Church, M.A., McLean, D.G. and Wolcott, J.F. 1987. River bed gravels: sampling and analysis. In: Thorne, C.R., Bathurst, J.C. and Hey, R.D., eds., *Sediment Transport in Gravel Bed Rivers*, New York: John Wiley & Sons, pp. 43–79.

Cope, O.B. 1957. The choice of spawning sites by cutthroat trout. *Proceedings of Utah Academy of Science, Arts and Letters* 34: 73–79.

Diepenbroek, M., Bartholoma, A. and Ibbeken, H. 1992. How round is round? A new approach to the topic 'roundness' by Fourier grain shape analysis. *Sedimentology* 39: 411–422.

Dietrich, W.E. 1982. Settling velocity of natural particles. *Water Resources Research* 18(6): 1615–1626.

Dietrich, W.E., Kirchner, J.W., Ikeda, H. and Iseya, F. 1989. Sediment supply and development of coarse surface layer in gravel bedded rivers. *Nature* 340: 215–217.

Drevin, G.R. and Vincent, L. 2002. Granulometric determination of sedimentary rock particle roundness. In: Talbot, H. and Beare, R., eds., *Mathematical Morphology. Proceedings of VIth International Symposium – ISMM 2002, Sydney, 3-5 April 2002*, Collingwood, VIC: CSIRO Publishing, pp. 315–325.

Dugdale, S.J., Carbonneau, P.E. and Campbell, D. 2010. Aerial photo-sieving of exposed gravel bars for the rapid calibration of airborne grain size maps. *Earth Surface Processes and Landforms* 35: 627–639.

Dunne, T. and Leopold, L.B. 1978. *Water in Environmental Planning*, New York: W.H. Freeman, 666pp.

Everest, F.L., McLemore, C.E. and Ward, J.F. 1980. *An Improved Tri-tube Cryogenic Gravel Sampler, Research Note PNW-350*, Portland, OR: US Department of Agriculture, Forest Service, Pacific Northwest Forest and Range Experiment Station.

Ferguson, R.I. and Paola, C. 1997. Bias and precision of percentiles of bulk grain size distributions. *Earth Surface Processes and Landforms* 22: 1061–1077.

Fripp, J.B. and Diplas, P. 1993. Surface sampling in gravel streams. *Journal of Hydraulic Engineering* 119: 473–490.

Gomez, B. 1984. Typology of segregated (armor/paved) surfaces: some comments. *Earth Surface Processes and Landforms* 9: 19–24.

Graham, D.J., Reid, I. and Rice, S.P. 2005a. Automated sizing of coarse-grained sediments: image-processing procedures. *Mathematical Geology* 37: 1–28.

Graham, D.J., Rice, S.P. and Reid, I. 2005b. A transferable method for the automated grain sizing of river gravels. *Water Resources Research* 41(7): W07020.

Graham, D.J., Rollet, A.-J., Piégay, H. and Rice, S.P. 2010. Maximizing the accuracy of image-based surface sediment sampling techniques. *Water Resources Research* 46(2): W02508.

Graham, D.J., Rollet, A.-J., Rice, S.P. and Piégay, H. 2012. Conversions of surface grain-size samples collected and recorded using different procedures. *Journal of Hydraulic Engineering* 138(10): 839–849.

Greeley, J.R. 1932. The spawning habits of brook, brown and rainbow trout and the problem of egg predators. *Transactions of the American Fisheries Society* 62: 239–248.

Groot, C. and Margolis, L. 1991. *Pacific Salmon Life Histories*, Vancouver: University of British Columbia Press, 564pp.

Hazzard, A.S. 1932. Some phases of the life history of the eastern brook trout Salvelinus fontinalis (Mitchill). *Transactions of the American Fisheries Society* 62: 344–350.

Hey, R.D. and Thorne, C.R. 1983. Accuracy of surface samples from gravel bed material. *Journal of Hydraulic Engineering* 109: 842–851.

Hilton, S. and Lisle, T.E. 1993. *Measuring the Fraction of Pool Volume Filled with Fine Sediment, PSW Research Note PSW-RN-414*, Berkeley, CA: US Department of Agriculture, Forest Service, Pacific Southwest Research Station, 11pp.

Hunter, J.W. 1973. *A Discussion of Game Fish in the State of Washington as Related to Water Requirements*, Olympia, WA: Fishery Management Division, Washington State Department of Game, 66pp.

Ibbeken, H. 1983. Jointed source rock and fluvial gravels controlled by Rosin's law: a grain size study in Calabria, South Italy. *Journal of Sedimentary Petrology* 53: 1213–1231.

Ibbeken, H. and Schleyer, R. 1986. Photo-sieving: a method for grain-size analysis of coarse-grained, unconsolidated bedding surfaces. *Earth Surfaces Processes and Landforms* 11: 59–77.

Inman, D.L. 1952. Measures for describing the size distribution of sediments. *Journal of Sedimentary Petrology* 22: 125–145.

ISO. 1977. *Liquid Flow Measurements in Open Channels – Bed Material Sampling, ISO 4364-1977 (E)*, Geneva: International Organization for Standardization.

Kellerhals, R. and Bray, D.I. 1971. Sampling procedures for coarse fluvial sediments. *Journal of the Hydraulics Division, American Society of Civil Engineers* 97: 1165–1179.

Klingeman, P.C. and Emmett, W.W. 1982. Gravel bed-load transport processes. In: Hey, RD, Bathurst, J.C. and Thorne, C.R., eds., *Gravel Bed Rivers*, Chichester: John Wiley & Sons, pp. 141–179.

Kondolf, G.M. 1988. *Salmonid Spawning Gravels: a Geomorphic Perspective on Their Distribution, Size Modification by Spawning Fish, and Application of Criteria for Gravel Quality, PhD Thesis*, Johns Hopkins University, Baltimore, MD.

Kondolf, G.M. 1997. Application of the pebble count: reflections on purpose, method, and variants. *Journal of the American Water Resources Association* 33(1): 79–87.

Kondolf, G.M. 2000. Assessing salmonid spawning gravels. *Transactions of the American Fisheries Society* 129: 262–281.

Kondolf, G.M. and Adhikari, A. 2000. Weibull vs. lognormal distributions for fluvial gravels. *Journal of Sedimentary Research* 70(3): 456–460.

Kondolf, G.M. and Li, S. 1992. The pebble count technique for quantifying surface bed material in instream flow studies. *Rivers* 3: 80–87.

Kondolf, G.M. and Matthews, W.V.G. 1986. Transport of tracer gravels on a coastal California river. *Journal of Hydrology* 85: 265–280.

Kondolf, G. M. and M. G. Wolman. 1993. Modification of gravel size by spawning salmonids. *Water Resources Research* 29: 2265–2274.

Kondolf, G.M., Cook, S.S., Maddux, H.R. and Persons, W.R. 1989. Spawning gravels of rainbow trout in Glen and Grand Canyons, Arizona. *Journal of the Arizona–Nevada Academy of Science* 23: 19–28.

Kondolf, G.M., Sale, M.J. and Wolman, M.G. 1993. Modification of gravel size by spawning salmonids. *Water Resources Research* 29: 2265–2274.

Krumbein, W.C. 1934. Size frequency distribution of sediments. *Journal of Sedimentary Petrology* 4: 65–77.

Krumbein, W.C. 1941. Measurement and geologic significance of shape and roundness of sedimentary particles. *Journal of Sedimentary Petrology* 11: 64–72.

Krumbein, W.C. and Pettijohn, F.J. 1938. *Manual of Sedimentary Petrography*, New York: Appleton-Century, pp. 31–33.

Lejot, J., Piégay, H., Hunter, P.D., Moulin, B. and Gagnage, M. 2011. Utilisation de la télédétection pour la caractérisation des corridors fluviaux : exemples d'applications et enjeux actuels. *Géomorphologie: Relief, Processus, Environnement* 2: 37–52.

Leopold, L.B. 1970. An improved method for size distribution of stream-bed gravel. *Water Resources Research* 6: 1357–1366.

Lisle, T.E. 1989. Sediment transport and resulting deposition in spawning gravels, north coastal California. *Water Resources Research* 25(6): 1303–1319.

Lisle, T.E. 1995. Particle size variations between bed load and bed material in natural gravel bed channels. *Water Resources Research* 31(4): 1107–1118.

Lisle, T.E. and Eads, R.E. 1991. *Methods to Measure Sedimentation of Spawning Gravels, PSW Research Note PSW-411*, Berkeley, CA: US Department of Agriculture, Forest Service, Pacific Southwest Research Station.

Lisle, T.E. and Hilton, S. 1992. The volume of fine sediment in pools: an index of the supply of mobile sediment in stream channels, *Water Resources Bulletin*, 28(2): 371–383.

Lisle, T.E. and Hilton, S. 1999. Fine bed material on pools of natural gravel bed channels. *Water Resources Research* 35(4): 1291–1304.

Lisle, T.E. and Madej, M.A. 1992. Spatial variation in armouring in a channel with high sediment supply. In: Billi, P., Hey, R.D., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel-bed Rivers*, Chichester: John Wiley & Sons, pp. 277–291.

Lisle, T.E., Nelson, J.M., Pitlick, J., Madej, M.A. and Barkett, B.L. 2000. Variability of bed mobility in natural gravel-bed channels and adjustments to sediment load at the local and reach scales. *Water Resources Research* 36(12): 3743–3756.

Lotspeich, F.B. and Everest, F.H. 1981. *A New Method for Reporting and Interpreting Textural Composition of Spawning Gravel, Research Note PNW-369*, Portland, OR: US Department of Agriculture, Forest Service, Pacific Northwest Forest and Range Experiment Station.

Marcus, W.A., Ladd, S.C., Stoughton, J.A. and Stock, J.W. 1995. Pebble counts and the role of user-dependent bias in documenting sediment size distributions. *Water Resources Research* 31(10): 2625–2631.

McNeil, W.J. and Ahnell, W.H. 1964. *Success of Pink Salmon Spawning Relative to Size of Spawning Bed Materials, Special Scientific Report – Fisheries 469*, Washington, DC: US Fish and Wildlife Service.

Meehan, W.R. 1991. *Influences of Forest and Rangeland Management on Salmonid Fishes and their Habitats, American Fisheries Society Special Publication 19*, Bethesda, MD: American Fisheries Society.

Milhous, R.T., Hogan, S.A., Abt, S.R. and Watson, C.C. 1995. Sampling river-bed material: the barrel sampler. *Rivers* 5: 239–249.

Mosley, M.P. and Tinsdale, D.S. 1985. Sediment variability and bed material sampling in gravel bed rivers. *Earth Surface Procedures and Landforms* 10: 465–482.

Otto, G.H. 1939. A modified logarithmic probability graph for interpretation of mechanical analyses of sediments. *Journal of Sedimentary Petrology* 9: 62–76.

Parker, G. 1990. Surface-based bedload transport relation for gravel rivers. *Journal of Hydraulic Research* 28(4): 417–436.

Parker, G. and Andrews, E.D. 1985. Sorting of bed load sediment by flow in meander bends. *Water Resources Research* 21(9): 1361–1373.

Parker, G. and Klingeman, P.C. 1982. On why gravel bed streams are paved. *Water Resources Research* 18: 1409–1423.

Pégot-Augier, T. 2012. *Caractérisation du Patron Granulométrique du Lit du Rhône par Imagerie, unpublished report, Mémoire de Fin d'Etudes d'Ingénieur CPE, UMR 5600 EVS*, University of Lyon, 56pp.

Petrie, J. and Diplas, P. 2000. Statistical approach to sediment sampling accuracy. *Water Resources Research* 36(2): 597–606.

Pettijohn, F.J. 1975. *Sedimentary Rocks*, 3rd edn., New York: Harper and Row, 628pp.

Pissart, A., Duchesne, F. and Vanbrabant. C. 1998. La détermination pratique des intervalles de confiance des comptages de cailloux et des mesures d'émoussé. Comparaison des mesures d'émoussé de Cailleux et de Krumbein. *Géomorphologie: Relief, Processus, Environnement* 3: 195–214.

Platts, W.S. and Penton, V.E. 1980. *A New Freezing Technique for Sampling Salmonid Redds, Research Paper INT-248*, Ogden, UT: US Department of Agriculture, Forest Service, Intermountain Forest and Range Research Station, 22pp.

Platts, W.S., Megahan, W.F. and Minshall, G.W. 1983. *Methods for Evaluating Stream, Riparian, and Biotic Conditions, General Technical Report INT-138*, Ogden, UT: US Department of Agriculture, Forest Service, Intermountain Forest and Range Research Station, 70pp.

Rice, S. and Church, M. 1996. Sampling surficial fluvial gravels: the precision of size distribution percentile estimates, *Journal of Sedimentary Research* 66(3): 654–665.

Rollet, A.J., Piégay, H., Dufour, S., Bornette, G. and Persat, H. 2014. Assessment of consequences of sediment deficit on a gravel river bed downstream of dams in restoration perspectives: application of a multicriteria, hierarchical and spatially explicit diagnosis. *River Research and Applications* 30(8): 939–953

Rood, K. and Church, M. 1994. Modified freeze-core technique for sampling the permanently-wetted streambed. *North American Journal of Fisheries Management* 14: 852–861.

Rosgen, D.L. 1996. *Applied River Morphology*. Pagosa Springs, CO: Wildland Hydrology.

Roussillon, T., Piégay, H., Sivignon, I., Tougne, L. and Lavigne, F. 2009. Automatic computation of pebble roundness using digital imagery and discrete geometry. *Computers & Geosciences* 35: 1992–2000.

Rubin, D.M. 2004. A simple autocorrelation algorithm for determining grain size from digital images of sediment. *Journal of Sedimentary Research* 74: 160–165.

Seal, R. and Paola, C. 1995. Observations of downstream fining on the North Fork Toutle River near Mount Saint Helens, Washington. *Water Resources Research* 31: 1409–1419.

Shirazi, M.A. and Seim, W.K. 1981. Stream system evaluation with emphasis on spawning habitat for salmonids. *Water Resources Research* 17: 592–594.

Shirazi, M.A. and Seim, W.K. 1982. Reply [to "Comment on 'Stream system evaluation with emphasis on spawning habitat for salmonids' by Mostafa A. Shirazi and Wayne K. Seim"]. *Water Resources Research* 18: 1296–1298.

Shirazi, M.A., Seim, W.K. and Lewis, D.H. 1981. Characterization of spawning gravel and stream system evaluation. In: *Proceedings of a Conference on Salmon-spawning Gravel: a Renewable Resource in the Pacific Northwest? Report 39*, Pullman, WA: Washington State University, State of Washington Water Research Center, pp. 227–278.

Tappel, P.D. and Bjornn, T.C. 1983. A new method of relating size of spawning gravel to salmonid embryo survival. *North American Journal of Fisheries Management* 3: 123–135.

Trask, P.D. 1932. *Origin and Environment of Source Sediments of Petroleum*, Houston, TX: Gulf Publications (as cited by Inman 1952).

Tukey, J.W. 1977. *Exploratory Data Analysis*, Reading, MA: Addison-Wesley.

USFWS. 1999. *Trinity River Flow Evaluation. Final Report: a Report to the Secretary of the Interior*, Arcata, CA: US Fish and Wildlife Service, Arcata Fish and Wildlife Office, and Hoopa, CA: Hoopa Valley Tribe, 307pp. plus appendices.

Vanoni, V.A. 1975. *Sedimentation Engineering*, New York: American Society of Civil Engineers.

Verdu, J.M., Batalla, R.J. and Martınez-Casasnovas, J.A. 2005. High-resolution grain size characterisation of gravel bars using imagery analysis and geo-statistics, *Geomorphology* 72(1–4): 73–93.

Wadell, H. 1932. Volume, shape and roundness of rock particles. *Journal of Geology* 40: 443–451.

Warrick, J.A., Rubin, D.M., Ruggiero, P., Harney, J.N., Draut, A.E. and Buscombe, D. 2009. Cobble cam: grain-size measurements of sand to boulder from digital photographs and autocorrelation analyses, *Earth Surface Processes and Landforms* 34(13): 1811–1821.

Wentworth, C.K. 1926. Methods of mechanical analysis of sediments. *University of Iowa Studies in Natural History*, 11(11): 3–52.

Whiting, P.J. and King, J.G. 2003. Surface particle sizes on armoured gravel streambeds: effects of supply and hydraulics. *Earth Surface Processes and Landforms* 28(13): 1459–1472.

Wilcock, P.R., Kondolf, G.M., Barta, A.F., Matthews, W.V.G. and Shea, C.C. 1995. *Spawning Gravel Flushing During Trial Reservoir Releases on the Trinity River: Field Observations and Recommendations for Sediment Maintenance Flushing Flows, Report CEDR-05-95*, Berkeley, CA: Center for Environmental Design Research, University of California at Berkeley.

Wilcock, P.R., Kondolf, G.M., Matthews, W.V.G. and Barta, A.F. 1996a. Specification of sediment maintenance flows for a large gravel-bed river. *Water Resources Research* 32: 2911–2921.

Wilcock, P.R., Barta, A.F., Shea, C.C., Kondolf, G.M., Matthews, W.V.G. and Pitlick, J. 1996b. Observations of flow and sediment entrainment on a large gravel-bed river. *Water Resources Research* 32(9): 2897–2909.

Williams, H., Turner, F.J. and Gilbert, C.M. 1982. *Petrography: an Introduction to the Study of Rocks in Thin Sections*, 2nd edn., San Francisco, CA: W.H. Freeman.

Wohl, E.E., Anthony, D.J., Madsen, S.W. and Thompson, D.M. 1996. A comparison of surface sampling methods for coarse fluvial sediments. *Water Resources Research* 32(10): 3219–3226.

Wolman, M.G. 1954. A method of sampling coarse river-bed material. *Transactions, American Geophysical Union* 35: 951–956.

Wolman, M.G. and Schick, A.P. 1967. Effects of construction on fluvial sediment, urban and suburban areas of Maryland. *Water Resources Research* 3: 451–464.

Yalin, M.S. 1977. *Mechanics of Sediment Transport*. Oxford: Pergamon Press, 298pp.

Young, M.K., Hubert, W.A. and Wesche, T.A. 1991. Biases associated with four stream substrate samplers. *Canadian Journal of Fisheries and Aquatic Sciences* 48: 1882–1886.

Zingg, T. 1935. Beiträge zur Schotteranalyse. *Schweizerische Mineralogische und Petrographische Mitteilungen* 15: 39–140.

CHAPTER 14

Coarse particle tracing in fluvial geomorphology

Marwan A. Hassan[1] and André G. Roy[2]
[1]University of British Columbia, Vancouver, BC, Canada
[2]Concordia University, Montréal, QC, Canada

14.1 Introduction

In this chapter, we present a review of sediment tracing techniques in gravel and sand bed rivers. The primary objective is to assess critically the use of tracers to obtain quantitative information of bedload sediment transport. Fine sediment tracing is covered in Chapter 9. Since most of the tracing techniques are site specific and to avoid tedious repetition, each method is briefly described. For more technical information, the reader should refer to the original material. Although we are aware that the topic has been reviewed in many countries, we mainly provide a summary of material either originally in, or translated into, English. After reviewing each tracing method, we present an evaluation of the techniques available for studying sediment transport in a fluvial environment.

Although considerable attention has been paid to specific influences that are known to determine river sediment transport rates and patterns, there is still a large inconsistency between data collected in the field and the results of theoretical and empirical models. This inconsistency is due to the large number of interrelated variables that affect sediment transport in rivers and also to the difficulties involved in field measurements. The numerous different field techniques have evolved to meet the complexities associated with understanding the processes of fluvial transport. A broad division can be made between, on the one hand, the use of traps and samplers in which moving sediment is collected during a flood event and, on the other, tracing the movement of individual grains between and during floods.

Characteristics of sediment movement would be much easier to obtain if it were possible to trace the movement of either the entire bed material or a given mass of individual grains. However, given the inherent difficulties in field sampling, this is not possible. Therefore, tracers are particularly valuable to fluvial geomorphology as they provide a means of overcoming technical and sampling problems. Tracers are defined as marked particles that are introduced into streams in order to obtain general information on the movement of sediment. Such labelling must permit traced sediment to be detected, and also operate within the fluvial environment in the same way as the natural material. Tracers provide an opportunity to study the general characteristics of sediment movement under varying flow conditions without the need for a detailed kinematic study of the sedimentary regime (Crickmore *et al.* 1990). They reveal the long-term action of the fluvial system and allow for the documentation of the interactions between sediment supply and channel morphology. Therefore, naturally or artificially labelled sediments are used within channel reaches to provide information on the rate and direction of sediment transportation, particle entrainment, periods of rest and movement of particles, step length of individual particles, residence time, flow competence, virtual rate of sediment movement, relations between distance of movement and flow strength, effect of physical characteristics on distance of travel, downstream fining, depth of the active layer, impact of particle sedimentological environment on distance of movement, sediment sources and depositional areas, volume of mobile sediment and wearing rate.

There have been reports of bedload tracing programmes in the field and laboratory since the late 1930s. To our knowledge, Einstein (1937) was the first to use tracers in a flume, and Takayama (1965) and Leopold *et al.* (1966) were pioneers in using painted tracers in the field. Since then, tracing natural clasts has become a common technique in many studies of river sediment transport. The most common and simplest method is to paint a clast so that it stands out from the rest of the bed material. However, the data are limited to material moving on the bed surface and hence the recovery rates of painted stones are extremely variable (Hassan *et al.* 1984; Hassan and Church 1992) and usually low (about 20%). Most studies (e.g. Leopold *et al.* 1966; Church 1972; Schick and Sharon 1974; Ashworth 1987) have emphasized the lack of simple relations between distance of travel and particle size. Since the late 1950s, new tracer techniques have been developed for both field and laboratory studies.

General overview of coarse particle tracing techniques

There are two general types of tracer techniques. One type uses passive tracers, which must be seen or sensed by an observer or a detector. The other type uses active tracers, which emit waves or rays detected by a spectrometer or receiver. Each approach has advantages and disadvantages (Table 14.1), including different recovery rates (Table 14.2). Passive tracers include exotic

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

Table 14.1 Summary of advantages, disadvantages and key research questions of various tracer methods.

Method	Advantages	Disadvantages	Lifespan	Key questions
Exotic (material and minerals)	Cheap, easy to apply, visual inspection of the surface	Limited to the surface, low recovery rate	Permanent	Downstream fining, wearing rate, sources and destinations, spatial dispersion
Painting	Cheap, easy to apply, visual inspection of the surface	Low recovery rate, limited to the surface, paint abrades after a few events	Depending on abrasion, typically <3 years	Travel distance, flow competence, sources and destination, virtual velocity, spatial dispersion
Fluorescent	Non-toxic, simple to inject, cheap, large quantity can be traced, useful for measurements on slowly evolving systems	Difficult to detect, limited to quantitative information, repeated studies are difficult in the case of long-life dye, wears out quickly, adheres to untagged particles, affected by temperature, light and salinity	Depending on dye type	Travel distance, flow competence, sources and destination, virtual velocity, spatial dispersion
Radioactive	High recovery rate, very powerful, detects both surface and buried particles	Toxic, hazard to environment and public, difficult to handle and inject, expensive, needs special laboratory and field equipment, limited to areas with no radioactive background, licensing constraints	Limited by half-life of material	Travel distance, flow competence, entrainment, virtual velocity, burial depth, 3D dispersion, volume of mobile sediment, sources and destinations
Iron oxide coating	Easy to apply, cheap	Limited to particles >11 mm, affected by noise from scrap metal, recovery may substantially disturb the bed, labour intensive, difficult to apply in deep water, coating abrades after a few events	As in paint	Travel distance, virtual velocity, depth or burial, 3D dispersion, flow competence, sources and destinations, volume of mobile sediment
Metal strips or plugs	Easy to apply, cheap	Affected by background noise, collar separation from pebbles, limited to large panicles, recovery may substantially disturb the bed, labour intensive, difficult to apply in deep water	Depending on abrasion/breaking of stones	As in iron oxide coating
Iron core	Long time expectancy, easy, cheap, can locate both buried and surface particles	Limited to large particles, low recovery rate, affected by iron mineral and scrap metal in the background, labour intensive, difficult to apply in deep water, might alter particle density, recovery may substantially disturb the bed	Permanent	Travel distance, virtual velocity, depth of burial, 3D dispersion, downstream fining, wearing rate, flow competence, sources and destinations, volume or mobile sediment
Inserted magnets	High recovery rate, can locate both buried and surface particles, cheap, easy to apply	Labour intensive, limited to sizes >11 mm, affected by background noise, difficult to apply in deep water, recovery may substantially disturb the bed	Permanent	As in iron core
Natural magnetic	No cost for tracers, unlimited number of tracers, easy to apply	Difficult to trace individual particles, detection system is fixed, needs field equipment and well-trained technician, affected by background noise, recovery may substantially disturb the bed	Permanent	As in iron core
Artificial magnetic	High recovery rate, easy to apply	Moderately expensive, affected by background noise, recovery may substantially disturb the bed, physical particle characteristics are different from natural material	Permanent	As in iron core
Magnetic enhance-ment	All sizes can be tagged, reasonable recovery rate, locate both buried and exposed particles	Cost of baking, requires furnace, high iron content in material, recovery may substantially disturb the bed, cracking under thermal stress, labour intensive, moderately expensive	Permanent	As in iron core
Radio transmit-ters	Very powerful, high recovery rate, can locate both buried and surface particles, continuous information on particle position during flood	Expensive, limited to large material, limited number of tracers	~1 year	Step length, rest duration, entrainment, travel distance, burial depth, particle trajectory, flow competence, sources and destinations, volume of mobile sediment, virtual velocity
PIT tags (RFID techniques)	Track individual particles, mobile and fixed detection systems high recovery rate, can locate both buried and surface particles, easy to apply, durability	Labour intensive, limited to sizes >40 mm, affected by background noise, difficult to apply in deep water, recovery may substantially disturb the bed, efficiency decreases with river size	Long (>50 years)	As in iron core

Table 14.2 Recovery rates of several bedload tracing projects.

Method	Reference	Tracers size range (mm)	Site	Recovery rate (%)
Exotic	Mosley (1978)	8–300	Tamaki River	5
	Kondolf and Matthews (1986)		Carmel River	
Painting	Einstein (1937)	17–24	Flume	High
	Takayama (1965)	22–128	Hayakawa	10–23
	Takayama (1965)	22–128	Fukugawa	21–27
	Takayama (1965)	22–128	Okawa	32–40
	Leopold *et al.* (1966)	75–150	Arroyo de les Frijoles	0–88
	Keller (1970)	Pebbles–cobbles	Dry Creek	41–65
	Church (1972)	Pebbles–cobbles	Ekalugad Rivers	High
	Slaymaker (1972)	Cobbles	Nant Calefwr	85–100
	Slaymaker (1972)	Cobbles	Nant Y Grader 9A	85–100
	Slaymaker (1972)	Cobbles	Nant Y Grader 9B	85–100
	Schick and Sharon (1974)	32–512	Nahal Yael	2–57
	Laronne and Carson (1976)	4–256	Seales Brook	5
	Thorne and Lewin (1979)	Pebbles–cobbles	Severn	40–79
	Leopold and Emmett (1981)	47–91	White Clay Creek	High
	Hassan *et al.* (1984)	45–180	Nahal Hebron	31–34
	Ashworth (1987)[*]	24–238	Allt Dubhaig	30–96
	Ashworth (1987)[*]	24–171	Feshie River	40–84
	Ashworth (1987)[*]	35–200	Lyngsdalselva	26–80
	Petit (1987)	Pebbles	La Rulles	High
	Tacconi *et al.* (1990)[†]	16–128	Virginio Creek	5–9
	Carling (1992)[†]	15–130	Carl Beck	98
	Carling (1992)[*]	15–130	Great Eggleshope	98
	Sear (1992; 1996)	Pebbles–cobbles	North Tyne	35–100
	Thorne (1996)[‡]	Pebbles–cobbles	Swale	86
	DeVries (2000)	22–254	Ranging River	>90
	DeVries (2000)	22–256	Issaquah Creek	>90
	Lenzi (2004)	32–512	Rio Cordon	52–100
	Wong *et al.* (2007)	7.5	Flume	High
	McNamara *et al.* (2008)	30–270	North Slope	13
	Hill *et al.* (2010)	4–9.5	Flume	High
Radioactive	Ramette and Heuzel (1962)	25–75	Rhone	100
	Stelczer (1968, 1981)	8–34	Danube	100
	Michalik and Bartnik (1986)	2–25	Wisloka River	100
	Michalik and Bartnik (1986)	2–25	Dunajae River	100
Ferric coating	Nir (1964)	52–240	Nahal Zin	4
Metal strips	Butler (1977)	34–116	Horse Creek	35
Iron core	Hassan *et al.* (1984)	45–180	Nahal Hebron	31–34
	Schmidt and Ergenzinger (1992)	50–170	Lainbach	17–92
Magnetic: inserted and artificial	Froehlich (1982)	Pebbles and cobbles	Homerka	High
	Ergenzinger and Conrady (1982)	Cobbles	Buonamico	100
	Hassan *et al.* (1984)	45–180	Nahal Hebron	90–93
	Reid *et al.* (1984)	29	Turky Brook	100
	Hassan (1990)	45–180	Nahal Og	55–56
	Hassan and Church (1992)	16–512	Harris Creek	75
	Hassan and Church (1992)	16–180	Carnation Creek	80
	Laronne and Duncan (1992)			75
	Lekach (1992)	45–180	Nahal Yael	100
	Schmidt and Ergenzinger (1992)[§]	60–137	Lainbach	25–100
	Hassan *et al.* (1995; 1999)	18–90	Metsemotlhaba	22–28
	Haschenburger (1996, 2011a,2011b)	16–200	Carnation Creek	>80[¶]
	Wathen *et al.* (1997)[‖]	23–362	Allt Dubhaig	50–100
	Stott and Sawyer (2000)	4–128	Tanllwyth	60
	Warburton and Demir (2000)	32–256	Lower Tees	89
	Warburton and Demir (2000)	32–256	Trout Beck	96
	Warburton and Demir (2000)	32–256	Upper Tees	100

Table 14.2 (*continued*)

Method	Reference	Tracers size range (mm)	Site	Recovery rate (%)
	Gottesfeld *et al.* (2004)	45–256	Forfar	60–100
	Gottesfeld *et al.* (2004)	45–256	O'Ne-ell	40–100
	Eaton *et al.* (2010)	32–90	Fishtarp	44–83
Enhanced magnetism	Arkell *et al.* (1983)	5.6–22.4	Plynlimon	63
	Sear (1992, 1996)	<22	North Tyne	5
Natural magnetism	Ergenzinger and Custer (1983)	>5	Squaw Creek	
Radio	Ergenzinger *et al.* (1989)[*] ∗	85–130	Lainbach	100
	Chacho *et al.* (1989)[†] †	60–100	Toklat	100
	Habersack (2001)	42–62	Waimakariri	100
	Lenzi (2004)	40–160	Rio Cordon	High
	McNamara and Borden (2004)	76–92	Reynolds	100
	McNamara *et al.* (2008)	50–150	North Slope	26
PIT tags	Nichols (2004)	57	Walnut Gulch	96
	Lamarre *et al.* (2005)	40–250	Moras	87–96
	Lamarre and Roy (2008a,b)	40–250	Spruce	57–92
	Rollet *et al.* (2008)	45–145	Ain	36
	MacVicar and Roy (2011)	40–280	Moras	66–94
	Bradley and Tucker (2012)	38–84	Halfmoon Creek	>93
	Liébault *et al.* (2012)	23–520	Bouinenc Torrent	25–78

Comments:
[a] See also Ashworth and Ferguson (1989).
[b] Reported in Hassan and Church (1992).
[c] Reported in Sear (1996).
[d] See also Gintz *et al.* (1996).
[e] For complete surveys.
[f] See also Ferguson *et al.* (1996) and Ferguson and Wathen (1998).
[g] See also Schmidt and Ergenzinger (1992) and Busskamp (1993, 1994).
[h] See also Emmett *et al.* (1990).
Based partially on Hassan *et al.* (1984), Hassan and Church (1992), Sear (1996) and Lamarre, *et al.* (2005).

and painted particles, fluorescent paint, radioactive elements, ferruginous, magnetic and transponders. The radio transmitter is the only available active tracing technique.

Both exotic and artificial tracers have been used to further our understanding of sediment transport in rivers. Given favourable circumstances, the mineralogical characteristics of exotic particles could serve or be tagged as natural tracers. However, their use is limited and there is a need to tag non-exotic sediment. In the latter case, artificial tracers are collected from a given experimental site and then processed in order to distinguish them from the remainder of the bed material. Each can be marked with an identification code specific to the individual clast before being reseeded back into the river. Comparing positions of clasts before, during and after a flow event can shed valuable light on the dispersion of individual clasts within the active bed layer.

There are aspects that are common to all tracer studies (Fig. 14.1). Each experiment entails defining objectives, selecting methods and materials that will effectively allow the realization of the objectives, determining sample size and recovery rate, injecting back into the river, detecting, data collecting, data analysing and interpreting. The successful use of tracers depends largely on the selection of methods and work

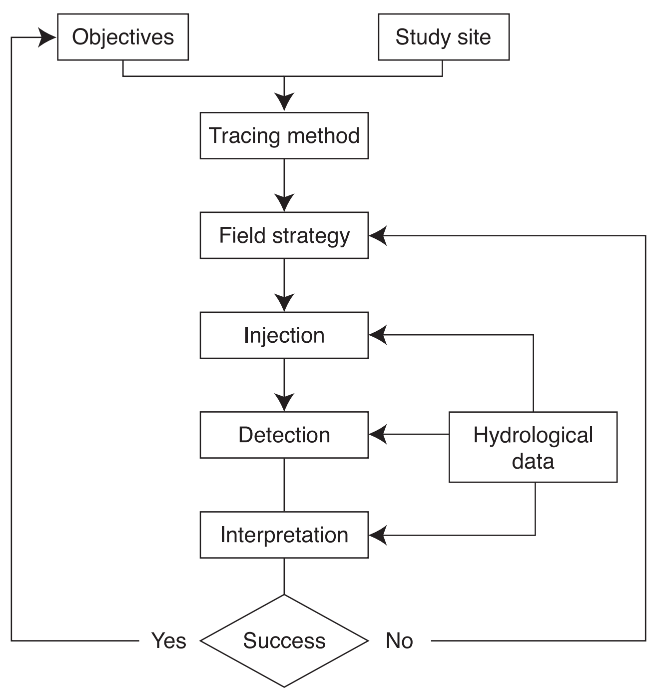


Figure 14.1 Flow chart of tracer experimental design.

procedures. Both study objectives and channel characteristics determine the type of tracers to be used, the duration of experiments and data analyses. If the objective is to study downstream sediment sorting, then the experiment should be carried on for a long time; therefore, a long-life tracer should be selected, such as magnetic tracers in gravel bed rivers. In some cases, where it is important to repeat an experiment in the same channel, a short half-life of fluorescent sand or radioactive tracers can be used. If the focus of the study is the behaviour of sediment within the active bed surface layer, then one should use magnetic, radioactive or passive integrated transponder (PIT) methods that are not limited to surface detection. In selecting a tracing method, channel size and morphology should be considered. Painted and magnetic tracers and PIT tags are suitable in small ephemeral channels or in shallow water where conditions promote a thorough search of the channel. In contrast, deep waters may limit the possibility of recovering magnetically or PIT tagged particles, so radioactive tracers or radio transmitters should be used.

Selecting a tagging technique also should ensure that the tracers are easily distinguishable but still accurately represent the natural sediment and its movement. This task is not easily achieved because the tracing technique may change some of the particle's physical characteristics, such as its density and hence its movement. In selecting a tagging technique, one should consider the suitability of the technique, including the presence of potential background noise, cost, recovery rate, detectability at low concentration levels, decay rate, ease of handling in both the laboratory and the field and safety for both the handler and the environment.

Recovery rate and population size

Very little is known about the population size needed to represent the movement of sediment in a stream channel. Researchers have rarely addressed this issue and hence the labelled population size has typically ranged from a few particles up to several hundred. Appropriate sample size depends on the study objectives and site characteristics. For example, reasonable estimates of mean travel distance may be obtained from a small sample of the order of a few tens of particles. However, in order to characterize the travel distance of individual particles, a sample of the order of 1000 is needed (Hassan and Church 1992). The same logic applies to channel size and length of the study reach. For small streams, the travel distance of individual particles may be obtained using a sample of the order of a few tens of tracers, whereas much larger samples are needed in large rivers (Hassan and Church 1992). However, the question of adequate sample size is still largely an open question in fluvial geomorphology and needs to be further studied.

All of the above points are correct provided that the recovery rate of the tracers is very high. However, the recovery rate depends on the tracing technique, sample texture, channel size, site characteristics and flow conditions. For techniques that are limited to detecting the particles at the bed surface (e.g. paint, exotic tracers), the recovery rate is usually low and depends on the texture of the tracers; it is low for small particles and high for coarse fractions (e.g. Leopold *et al.* 1966; Laronne and Carson 1976). In comparison, radioactive, magnetic and PIT techniques allow the detection of buried tracers and are therefore more likely to provide high recovery rates. In cases where the bed material contains natural magnetic particles, the recovery rate of magnetic particles may be low. For a given fluvial system, the recovery rate will also depend on the magnitude of flow events. Small flow events are likely to disperse tagged stones near the bed surface whereas larger flow events will mobilize the entire bed and are likely to result in deeper burial of particles. Deeply buried particles are more difficult to recover than near-surface particles because they may be beyond the sensitivity of the detection system. For example, the recovery depth of particles marked with PIT tags is typically around 25 cm, thus leaving undetected all particles buried deeper that this threshold.

Injection

Once the tagging technique has been selected and the tracers prepared, the manner in which the sediment is introduced into the fluvial system should be determined. This will vary from one experiment to another and depends mainly on the objectives and the physical conditions of the study site. In all cases, the tagged sediment should be introduced in a way that reduces the effects of artificial seeding and corresponds to the natural sediment transport conditions. Since this task is difficult to achieve, researchers treat the initial movement following seeding with caution (e.g. Hassan and Church 1992; Schmidt and Ergenzinger 1992; Ferguson and Hoey 2002). The volumetric method is one way to seed particles into a river. A trench across the width of the channel is excavated to the estimated depth of the active layer; the sediment is tagged and then returned to the trench (Hassan 1988; Wilcock *et al.* 1996a,b). However, a large proportion of the tagged sediment is likely to remain in place except in large flows (e.g. Hassan 1988). In the case of sand, the material is usually seeded at a point or across the channel width (e.g. Sayre and Hubbell 1965; Rathbun and Kennedy 1978). Several methods have been used for seeding pebble and cobble tracers:

1. seeding on the bed surface along lines across the channel (e.g. Leopold *et al.* 1966; Hassan *et al.* 1984, 1999; Roy and Bergeron 1990);
2. replacing a local stone found in the bed (e.g. Leopold and Emmett 1981; Ashworth and Ferguson 1989; Schmidt and Ergenzinger 1992);
3. selecting random particles, tagging and returning them to the same places (Hassan *et al.* 1991; Hassan 1993) or replacing a natural particle with a particle of exotic lithology of the same size and shape, placed into the hole from which the native particle was removed (Rovira and Kondolf 2008);
4. dumping tracers into a river (e.g. Mosley 1978);
5. volumetric placement in a trench (Hassan 1988; Wilcock *et al.* 1996a,b);
6. marking in situ (Ritter 1967).

To the extent that the positioning of the tagged particles on the bed surface is artificial and more exposed than natural positions, this will influence sediment movement during at least the first transporting flow event (e.g. Hassan *et al.* 1991; Ferguson and Hoey 2002). If carried out successfully, however, method **6** and possibly **3** and **2** above may avoid this problem.

Detection

The detection strategy and method depend largely on the study objectives and on the tagging technique. In selecting the detection method, one should aim that both buried and surface particles can be recovered. This will provide information on sediment dispersion within the active bed layer. Detection methods include visual identification, magnetic or metal detection, radioactive detection, neutron activation and radiofrequency wave detection. Some detection methods permit individual identification of tracer particles such as paint, magnets and radio [radiofrequency identification (RFID)-based methods], whereas some do not (e.g. radioactive material). This affects the type of recovered information and hence data analysis and interpretation. The following detection strategies have been used:

1. *Individual location of particles after a flow event:* The bed is searched between flow events, providing information on net particle movement during an entire flow event. Particles are then located visually or detected by metal, magnetic or radiofrequency sensors.
2. *Fixed automatic detection of tagged particles:* This system provides automatic detection of the passage of the tracers during a flow event. Three magnetic detection systems of this kind are described in the later subsection 'Magnets'. There is also the possibility of mounting fixed antennas to detect the passage of particles marked with PIT tags. A detecting device can be mounted on a boat or a low bridge.
3. *Sample and search:* Samples are taken from different parts of the bed. Particles are detected visually as in the case of fluorescent dye or by a sensor for radioactive or magnetic material.
4. *Periodic dynamic detection of the tagged particles:* The dispersion area of the tagged particles is surveyed during a transporting flow event. Such information can be used to prepare dispersal maps of the tagged particles based on time intervals during a flow event. To do so, the detection device should be mounted on a boat such that the entire dispersal area can be searched. Tracers are detected by scintillation detectors or radio transmitters.
5. *Automatic dynamic detection of tagged particles:* This method allows the continuous positioning of the tagged particles and provides very detailed information on their movement during a flow event. Radio transmitters are used for locating these tracers.

Interpretations

Data analyses and interpretations depend largely on the study objectives, experimental design and tracing method, all of which control the quality of the data. Qualitative and quantitative methods of data analysis have been used. There is some subjectivity in the qualitative analysis, which may contain considerable errors and lead to misinterpretations of data. In contrast, the objective of a quantitative analysis is to obtain information on the movement of sediment, such as direction of dominant movement, spatial dispersion, the relative mobility of sediment between events or different sites and estimates on the depth of the active bed layer.

Crickmore *et al.* (1990) described three quantitative methods for data interpretation and sediment discharge calculations: time integrated, steady dilution and spatial integration. Both the injection method and the study objectives determine what method to use for a particular case. In the time-integrated method, a single dose of tracers is introduced into the flow and then the concentration at a given downstream cross-section is monitored continuously. The distance between the injection point and the monitoring station should be long enough to allow sufficient mixing of sediment over the study site. In cases of slow-moving sediment such as gravel, the distance should be short but decreasing with particle size. The steady dilution method is not suitable for coarse bed material and therefore is not described here. Spatial integration is the most widely used method in sediment transport studies. Sediment discharge is obtained using the velocity of the tracers and the depth of the active layer. The velocity of the tracers is determined from the positional shift of the centroid of the spatial distribution as mapped at time intervals (Crickmore *et al.* 1990). Usually, sediment cores taken over the study area determine the mean depth of the active layer. The method is based on the assumption that the sediment transport is close to uniform over the study reach, an assumption that is difficult to satisfy in nature (DeVries 1973; Crickmore *et al.* 1990). Theoretically, this method is applicable to both suspended and bedload modes of sediment transport. This method has been used to calculate sediment transport in sand bed rivers (Sayre and Hubbell 1965) and gravel bed rivers (e.g. Hassan *et al.* 1992; Haschenburger and Church 1998). Sources of error in all stages of the experiment should be examined and resolved. In the case of artificial placement of tracers, data collected after the first flow event should be handled with caution. This problem could be overcome by in situ marking and to a certain extent by random selection of particles from the bed surface, returning them to the same positions.

Hydraulic data

Most tracer studies include the collection of hydraulic and geomorphic data from the study site. Such information is necessary for data interpretation and extrapolation. The hydraulic data are collected in order to establish relations between sediment transport characteristics and flow parameters. Appropriate hydraulic and geomorphic data depend largely on the study objectives and site characteristics. Hydraulic data that have been used include flow hydrograph, peak discharge, water depth, mean flow velocity, water surface slope, shear stress and stream power.

In some fluvial systems, for example desert streams, it is difficult to obtain detailed hydrological data and researchers use general hydraulic variables such as peak discharge. Tracer data such as travel distance, burial depth and virtual velocity have been related to hydraulic variables, including peak discharge, shear stress and stream power (e.g. Sayre and Hubbell 1965; Leopold *et al.* 1966; Reid *et al.* 1984; Hassan *et al.* 1992; Hassan and Church 1994; Haschenburger and Church 1998; Haschenburger 2011a,b).

Measured geomorphic characteristics include channel size, bedforms, bed texture, surface structures, bank stability and sediment supply from adjacent slopes and upstream tributaries. Both channel bedforms (e.g. bars, pools and riffles) and surface structures (e.g. pebble clusters, imbrications) impact particle entrainment probability and hence the distance of travel. Both the initial position before entrainment and the final position after movement should be described. The outcome of the study depends also on the amount of sediment supply from slopes and channel banks. Sediment supply to the channel will likely influence relations between flow parameters and movement characteristics.

14.2 Tracing methods

Exotic particles

The simplest form of tracing is probably the injection of exotic bed material into a river. Although the method is cheap and grain size representation can be excellent, the recovery rate of exotic particles is generally low because of visual identification of surface particles. Therefore, the use of this method to determine sediment transport is considered to be inadequate because of lack of information on the depth of the active layer. Furthermore, the use of the surface distribution of tracers to define travel displacement distribution, virtual velocity and volume of sediment transport is questionable (Hassan *et al.* 1991; Sear *et al.* 2000; Ferguson and Hoey 2002).

Mosley (1978) introduced a truckload (about 3 m^3) of limestone aggregate into the Tamaku River in New Zealand. The limestone aggregates had a golden colour, easily distinguishable from the greywacke transported by the river. The aggregates ranged in size from sand up to boulders, similar to the size range found in the river. After a mobilizing event, the channel bed was searched and the movement distance of particles larger than 8 mm was recorded. The recovery rate was low, approximately 5% of the volume.

Kondolf and Matthews (1986) used a similar method to trace the movement of white dolomite fragments introduced into the Carmel River, California, as rip-rap to combat bridge erosion. Since the rip-rap was introduced into the flow at a defined point, variation in the downstream distribution of tracers provided some useful information. As the introduction of tracer material into the flow was dependent on rip-rap erosion, the values of distances moved were artificially lowered by the newly introduced particles. The main problem faced when dealing with the introduction of exotic rock is whether the characteristics of the new material – density, shape, roundness and size distribution – are comparable to those of the natural material found in the channel (Table 14.1). In the study by Kondolf and Matthews (1986), the introduced material was more angular than the natural material, whereas the opposite was true in the study by Mosley (1978).

Natural labelling of sediment may be provided by the presence of mining waste, which could be divided into mineral and clastic tracers (e.g. Lewin and Macklin 1987; Knighton 1989; Macklin and Lewin 1989; Macklin, *et al.* 1992; Hattingh and Rust 1993; Langedal 1997; Houbrechts *et al.* 2011). The minerals, mostly sand or finer material, can be used for the long-term study of spatial dispersion and residence time in floodplains, channels and terraces, whereas clast sizes can be used in the channels and bars. When using such material, it is necessary to ensure that the physical characteristics of the mining material are comparable to those of the natural material found in the channel. Chapter 3 covers this topic.

Painted particles

Paint is a simple method of tracing bed material in a river. This method has the same disadvantages and advantages as the use of exotic particles. In this method, particles taken from the channel are painted so as to stand out against the rest of the bed material. Each particle is identified by a number. The entire tracer population can be painted one colour or colours may differ according to class size or point and time of insertion or other conventions. After each flow, the entire length of the study area is searched for painted particles. Those found on the bed surface are recorded, their morphological and sedimentological environments are described and they are then replaced on the bed surface to await the next flood.

Leopold *et al.* (1966) found a correlation between the strength of flow events and the recovery rates of painted tracers. The relation between clast size and recovery was further established by Laronne and Carson (1976), who found that the recovery rate ranged between 100% for large particles and as low as 0.5% for the smallest. In spite of these limitations, this low-cost method is still in use in both laboratory (Wong *et al.* 2007; Hill *et al.* 2010) and field (McNamara *et al.* 2008) settings.

Fluorescent paint

In fluorescent dye tagging, an amount of sediment is labelled with a dye which, upon stimulation by light of a suitable wavelength, emits light of a wavelength characteristic of the dye. Techniques for coating grains have not been standardized and several methods can be found in the literature. Ingle (1966) gave a detailed description of coating techniques. The fluorescent method is suitable for sand tracing, but can also be used for larger material (Yano *et al.* 1969). Sediment has to be washed, thoroughly air dried and then sieved into size fractions, each of which can be painted a different colour to yield insights into differential mobility. Fluorescent dyes suitable for sediment tracing experiments include rhodamine (red), auramine (yellow), eosine (green–yellow), primulin (dark blue), fluorescein (green) and anthracine (blue–violet) (Shteinman *et al.* 1997).

To create the fluorescent tags, a given amount of sediment is placed in a motor-driven cement mixer (e.g. Kennedy and Kouba 1970; Rathbun *et al.* 1971). One of either acetone, agar, ethanol or chloroform solution is then added with a small amount of resin to the mixture. The selection of the solution type depends on the dye characteristics and the desired life expectancy of the tracer. For example, agar lasts for a few days whereas chloroform can last for a few weeks (Shteinman *et al.* 1997). As the grains tumble, they are coated with a thin layer of solution. The dye solution should be added until fluorescence of the material reaches an adequate level. The mixture should then be spread out on a polyethylene sheet to dry and for the solution to evaporate. This work should be conducted outside, in a strong breeze. The resulting coated aggregates are placed between rubber rollers, or other similar device, to separate them into single grains. In order to avoid the tendency for the coated particles to cluster, one can add detergent to the material before injection into the river.

After each flow event, core samples are taken along the study reach. The core samples permit one to assess the vertical and downstream dispersion of the traced material. The collected samples should be air dried, examined under ultraviolet light to determine the presence of the traced material and tracers counted by their number per unit weight of the bulk material. To represent the spatial dispersion of the tagged particles, the collection of a large number of samples is recommended. Visual analysis of the samples can be laborious and hence expensive. Some instruments have been developed for automatic counting of the tagged particles but their efficiency is not clear (for more information, see Nelson and Coakley 1974; Coakley and Long 1990).

Radioactive tracers

Radioactive tracing was first used in the late 1950s and early 1960s (e.g. Hours and Jeffry 1959; Lean and Crickmore 1960; Hubbell and Sayre 1964; Sayre and Hubbell 1965). The presence of radioactively labelled sediment is determined by detecting the radiation given off by the tag. A wide range of radioisotopes allows a choice of tracers with half-lives that match the study objectives. For example, radioactive tagging with a short half-life can be used for repetitive studies on the same fluvial system.

Processes for attaching radioisotope tags to sediment particles have been reviewed in detail by Ariman *et al.* (1960), Petersen (1960) and Caillot (1970, 1983). The radioactive tracers can be grouped into three categories according to the method of labelling. The first involves the use of manufactured radioactive grains of glass of the same density and shape as the bed material. The second consists in applying the isotope to the outside or within the natural grains. In the outside application technique, the isotope is incorporated into a glue, forming a thin layer on the particle surface. This approach has been widely used in studies of bedload transport in gravel bed rivers. The third method involves the introduction of a natural radioactive mineral into the fluvial system.

The most commonly used isotopes are ^{46}Sc (half-life 83.9 days), ^{110}Ag (half-life 253 days), ^{51}Cr (half-life 27.8 days) and ^{140}La (half-life 40.2 hours). Isotopes such as ^{60}Co are considered to be dangerous because of their long half-life (Nelson and Coakley 1974). For safety reasons, it is recommended that the radioactive material be introduced into an isolated section of the river. A crystal or plastic scintillator optically coupled to a photomultiplier tube can be used to detect the radiation emitted by the tagged material.

In order to study the movement of coarse particles, the isotope is introduced into the particles themselves. A small hole is drilled into the particle, then the isotope is inserted and the hole is sealed with cement. Since the movement of coarse bedload is very sporadic and slow, long-lived isotopes may be appropriate. This method has been used in several studies concerned with the movement of coarse material and has proved to be successful (Ramette and Heuzel 1962; Stelczer 1968, 1981; Michalik and Bartnik 1986, 1994). However, the procedure is so time consuming that the total number of tagged particles is limited to a few tens per site.

According to Crickmore *et al.* (1990), the radioactive technique is the most versatile tracing method that can be used in silt, sand and gravels. Owing to its toxic nature, however, this method has not proved as popular as first expected (Crickmore *et al.* 1990). Radioactive tracers allow in situ, continuous detection of exposed and buried particles and make possible immediate data processing and hence evaluation of the tracking strategy (Crickmore *et al.* 1990). They also offer the opportunity for detailed mapping of the tracers over large areas and a range of time intervals. The main disadvantages are their toxicity and the hazard they pose to the environment, difficulties in handling and their expense.

Ferruginous tracers

Iron oxide coating was used in one of the first attempts made to locate buried clasts. Nir (1964) painted synthetic concrete cobbles with an iron oxide coating. As shown in Table 14.2, recovery rates were low, primarily due to the low sensitivity of the detection equipment. However, with modern detectors, this problem can be overcome. Butler (1977) tagged particles for relocation with a metal detector by wrapping strips of aluminium around them. However, attachment was a problem and several of the aluminium collars were found to have broken away from their pebbles, contributing to the large loss rates reported in Table 14.2. In addition, the presence of iron minerals in the background rock seriously interfered with the signals received by the detector. The preparation and use of iron tracers are the same as described below for magnetic tracers. The method is effective in locating shallow buried particles down to a few centimetres using a metal detector (Schmidt and Ergenzinger

1992). In order to increase the recovery rate, iron pieces of the order of 3 cm are recommended. However, this limits the size of the particles that can be used and might alter the particle density (Bunte and Ergenzinger 1989). With modern detectors, however, small pieces of iron can be detected.

Magnets

Magnets can be inserted into pre-drilled holes situated at the centre of gravity of each particle. The magnet used in most studies is ceramic and can be manufactured in different sizes and shapes. The magnetic field is unaffected by changes of the environment, weather or matrix. In order to increase the recovery rate, the size of the inserted magnet should be as large as possible; however, it is limited by the particle size and density. A large magnet is apt to alter the particle density and change its behaviour relative to natural ones. After insertion of the magnets, the cavity in each particle is filled with transparent epoxy. To identify particles, numbers can be inserted just inside the cavity and then covered again by the transparent epoxy. The drilling and the magnet insertion should not affect the particle strength or significantly alter its density (Hassan *et al.* 1984). Magnetic locators can be used to find the tagged stones. This method permits the location of buried particles up to 1 m underneath the bed surface, as well as on the surface (e.g. Hassan *et al.* 1995; Gintz *et al.* 1996; Eaton *et al.* 2010; Haschenburger 2011a,b). It is, however, a very tedious and time-consuming job to dig and record the tagged particles and it disturbs the channel bed extensively.

Natural sediment with a high magnetic content can be also used as a tracer. Ergenzinger and co-workers used natural magnetic cobbles and pebbles in Squaw Creek to estimate bedload transport rates during floods. The subsection 'Automatic detection of magnetic tracers' below gives a description of this study.

If no naturally magnetic material is present where studies are to be made, it is necessary to use artificially magnetized tracers. Reid *et al.* (1984) manufactured synthetic magnetic clasts made from resin and crushed barites with a ferrite rod core. These would make ideal classical tracers if it were not for the expense and the uniformity of shape and size when produced in bulk from moulds. The motion of the particles during a flood was registered by their passage over a sensor that distorts the magnetic field, causing a change in the inductance of the coils installed across the river bed. A metal detector was used to locate the tracers after a flow event.

To examine the impact of particle size and shape on travel distance and burial depth, Schmidt and Ergenzinger (1992), Gintz (1994) and Gintz *et al.* (1996) used coloured concrete. They manufactured artificial cobbles of about the same weight and density as natural sediment, but in different shapes. The concrete was formed in moulds with a magnetic core in the centre. Owing to difficulties with drilling and inserting magnets in small particles, Hassan *et al.* (1995) and Hassan and Church (1992) manufactured small stones made of resin, magnets and small pieces of lead shot, to adjust the overall density. Using the artificial stones, they were able to tag and release clasts as small as 8 mm.

Using the magnetic tagging method, Hassan *et al.* (1991) (see also Schick *et al.* 1987) examined the three-dimensional (3D) dispersion of coarse material and its relation to the mechanism of the scour layer in Nahal Hebron, an ephemeral stream in the Negev Desert. Of 282 tagged particles, 66% were found on the bed surface and 34% were buried (Fig. 14.2a). Figure 14.2(a) demonstrates, flood by flood, a clear pattern of sediment exchange between exposed and buried particles. The vertical

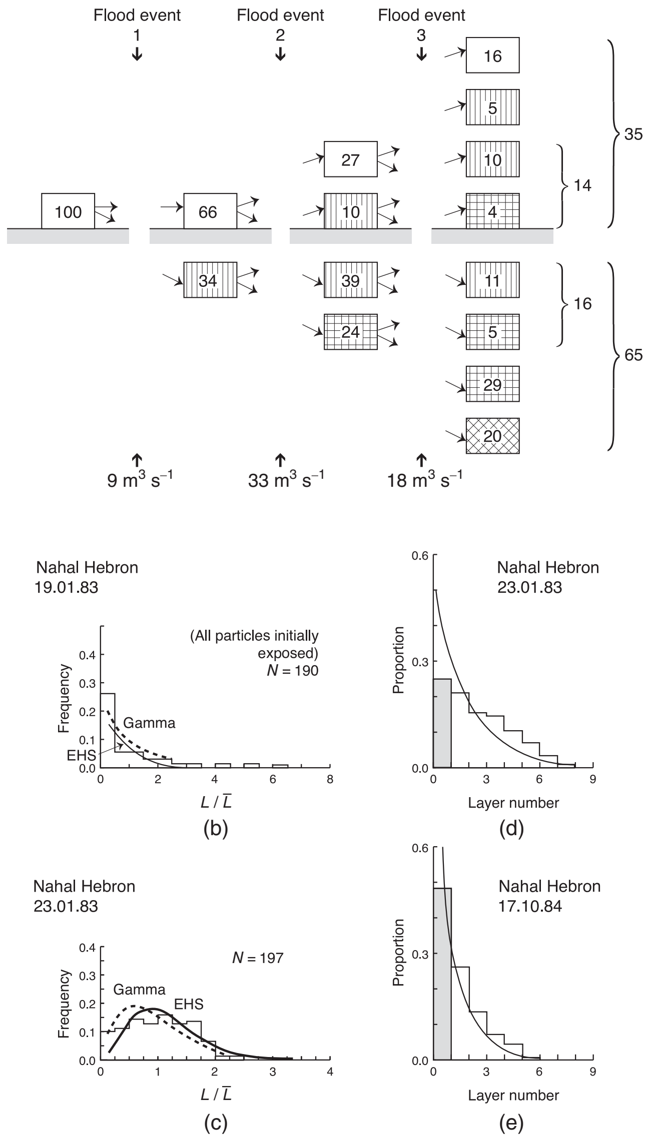


Figure 14.2 (a) Vertical exchange of tagged particles within the active layer as a result of flow events in Nahal Hebron. Boxes denote sediment exchange between the surface and subsurface and do not represent burial depth. Source: Shick *et al*, 1987. Reproduced with permission of Geological Society of London. (b, c) Travel distance distribution of all tagged particles in Nahal Hebron. EHS is the Einstein–Hubbell–Sayre distribution, L/L represents the scale distance of movement and layer number is the scaled burial depth. Source: Hassan *et al*, 1991. Reproduced with permission of AGU. (d, e) Burial depth distribution of tagged particles in Nahal Hebron. Source: Hassan and Church, 1994. Reproduced with permission of AGU.

exchange of the particles is subject to local influences such as channel morphology (e.g. pools, riffles and bars), slope and shielding by other particles. Hassan (1990) related the depth of burial of the tagged particles to the depth of fill and scour in the study site. Figure l4.2(b) and (c) provide examples of travel distance distribution after a flow event in Nahal Hebron (Hassan *et al.* 1991). Based on data collected from Nahal Hebron and other rivers, bed surface structure appeared more important in controlling the travel distance than the particle size (Church and Hassan 1992; Hassan and Church 1992). The distribution of burial depth of tagged particles in Nahal Hebron fits the exponential function well (Fig. 14.2d and e) and serves as the basis for a model describing the vertical mixing of sediment (Hassan and Church 1994).

Artificial magnetic enhancement

It has been noticed that, after forest fires, the magnetic content of soil particles is enhanced to a level that can be detected (Rummery *et al.* 1979). This magnetic tracing method is based on the enhancement of natural magnetism by high-level heating of naturally iron-rich fluvial pebbles and their reintroduction into the stream bed for subsequent tracing. This phenomenon has also been used to trace the sources of suspended sediment in drainage basins, as shown in Chapter 9 (Oldfield *et al.* 1979; Walling *et al.* 1979). Oldfield *et al.* (1981) heated clasts to temperatures ranging between 200 and 1150 °C and found that 900 °C yielded optimum results in terms of magnetic enhancement. Major changes in the bulk density of the material were observed for temperatures greater than 1000 °C. The best results were obtained by rapid heating with the sample inserted in a preheated oven close to the peak temperature of 900 °C. The heating time ranges between 20 min for small particles and up to 2 h for large particles. Rapid cooling of the samples, in either air or water, gives rise to higher levels of magnetism (Oldfield *et al.* 1981). Through the heat treatment, the particle mineralogy is altered and the magnetism is enhanced up to 300 times its original power, a level that can be detected and distinguished from the bed material. The method has been used to trace bedload in small forest ditches in the Welsh uplands (Arkell *et al.* 1983) with a recovery rate of 63%. However, magnetic enhancement yielded very low recovery rates in the North Tyne River (Table 14.2) (Sear 1992, 1996), which can be attributed to the larger size of that system. This demonstrates the need to consider channel scale in selecting a suitable tracing method.

Automatic detection of magnetic tracers

Automatic detection systems can track the movement of natural, artificial or inserted magnetic tracers. The underlying principle is that when a magnet passes over an iron-cored coil of wire, a measurable electronic pulse is generated.

In the Buonamico River, Calabria, Ergenzinger and Conrady (1982) inserted magnets into holes drilled in pebbles and a magnetic detector was used to monitor their passage. A similar system was used to detect the passage of naturally magnetic cobbles and pebbles past a fixed point during flow events in Squaw Creek, Montana (Ergenzinger and Custer 1983). The Squaw Creek system (Fig. 14.3) consisted of four wire coils, 1.4 m apart, connected in series and protected from water with several layers of silicon. The wires were wrapped around a 1 m long, 2 cm diameter iron bar. Each coil consisted of 9000 windings of 0.2 mm copper wire. Hassan (1988) achieved similar detection results by using a pipe 10 cm in diameter with 3000 windings of 0.2 mm copper wire. The Squaw Creek system was bolted in a slotted concrete block 1.25 m long, 0.2 m wide and 0.15 m high. The slot was covered with aluminium sheet metal to protect the detector from the impact of the passing stones. The wires of the detectors were placed in copper tubing that ran under a log that had been installed across the channel width and were connected to an amplifier, filter and flat bed recorder in a shelter on the riverbank. In Squaw Creek, two detectors were installed below a log that forced the pebbles and cobbles to overpass the sensors. As a result of the smooth surface of

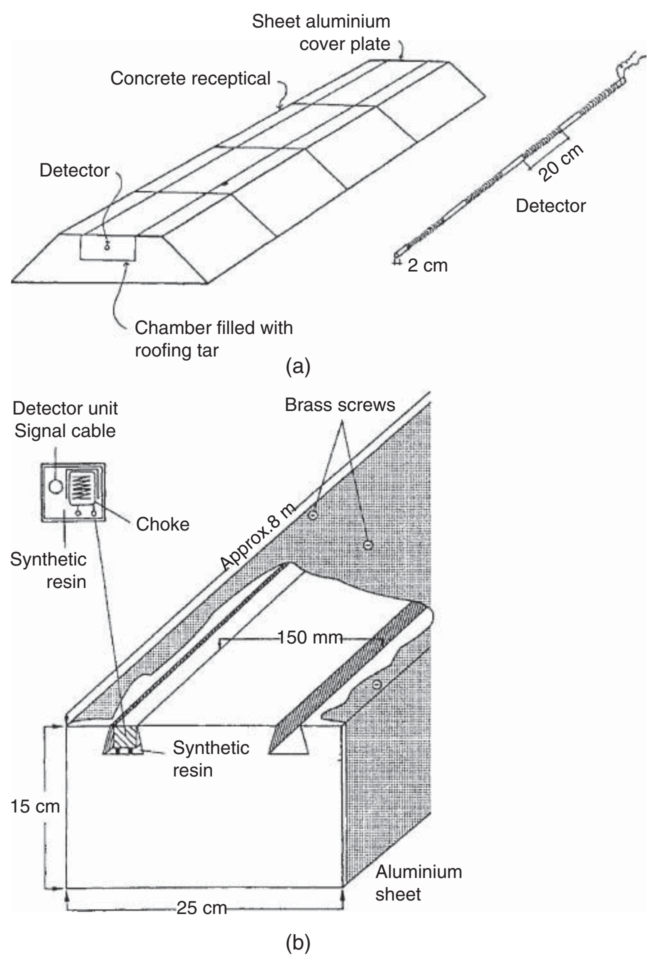


Figure 14.3 (a) Schematic diagram of the automatic magnetic detecting system. Source: Ergenzinger and Custer, 1983. Reproduced with permission of AGU. (b) Cross-section of the detector log. Source: Spieker and Ergenzinger, 1990. Reproduced with permission of IAHS.

the aluminium sheet metal and the positioning of the detector immediately below the log, no sediment accumulated on the system.

This type of system provides in situ continuous measurements of bedload movement during flow events and an unlimited number of stones can be recorded. However, it is an expensive method and requires a considerable knowledge of electronics. Furthermore, as the signals are a function of particle velocity, magnetic content and distance from the sensor, it is difficult to calibrate the signals and convert them to the number of stones. Using a sophisticated system of demodulation and electronic data processes, Bunte *et al.* (1987) and Spieker and Ergenzinger (1990) were able to detect stones as small as 3 cm in diameter.

Reid *et al.* (1984) used a commercially built system that worked in the same fashion as a metal detector. The system consisted of two elongated unscreened coils, each 2.3 m long. The sensors were fully balanced over the entire width of the channel. The passage of the tracers over the sensor distorted the magnetic field and produced a change in the inductance of the coils. The detected signal was amplified and demodulated to produce a change in voltage that was recorded on a chart. To avoid double registration of tracers and the influence of particles settling on or very close to the system, a self-balancing system that tuned out the influence of such particles after a predetermined time interval was built into the circuit. The system operated automatically and was activated by circuit closure in the mercury tilt-switch that was attached to the water stage recorder. Two sensors were installed, 11 m apart, on a straight reach in Turkey Brook, England. The main advantage of both systems is the automatic detection of the traced particles. In addition, the system of Reid *et al.* (1984) allows for individual detection of the tracers after a flow event. However, the system is expensive and is fixed in one position.

The most recent sophisticated magnetic system was developed by Tunnicliffe *et al.* (2000), known as the bedload magnetic detector (BMD) system. It consists of an array of sensors housed in an aluminium beam. Each sensor consists of a copper coil set inside a strong, vertically magnetized doughnut-shaped magnet. An iron casing serves as a Faraday cage to confine the magnetic field of the sensor so that it is not influenced by magnetic objects beyond the projected sensor area. Each sensor is digitally sampled via analogue–digital recorders. In the laboratory, Hassan *et al.* (2009) tested the BMD under controlled conditions in order to examine the influence of some variables on the performance of the system. Hassan *et al.* (2009) developed empirical relationships between the signal variables and passing object characteristics. The empirical relationships are based on the underlying theory of the electromotive force generated by a susceptible object passing through a magnetic field, but they vary from the theory because of the variable particle shape and non-ideal geometric configuration of the magnetic detectors. Results from the laboratory work showed that it is possible to relate signals from the BMD sensor to particle volume and speed. However, Hassan *et al.* (2009) identified major problems with the performance of the sensor. With the current sensor design, the strength of the magnetic field varies greatly from the edge to the centre of the sensor. This causes large differences in the signal response to the same particle passing over the edge or the centre of the sensor. Therefore, they suggested a new design that could overcome these problems and improve the performance of the system.

Radiofrequency identification and passive integrated transponders (PIT tags)

Recently, passive integrated transponders (PITs, commonly referred to as *PIT tags*) have been successfully used to track particles in rivers (Lamarre *et al.* 2005; Carré *et al.* 2007; Lamarre and Roy (2008a,b); Rollet *et al.* 2008; MacVicar and Roy 2011; Bradley and Tucker 2012), in soils (Wilson *et al.* 2010), at hillslope–channel transitions (Nichols 2004) or on beaches (Allan *et al.* 2006; Curtiss *et al.* 2009; Bertoni *et al.* 2010). Introduced as part of a security system, PIT tags have been widely used since the 1980s to study animal behaviour, habitat preferences, movement and migratory patterns. Several innovative applications have been developed in fish ecology.

PIT tags use the radiofrequency identification (RFID) principle that is based on the recovery of a signal from an electronic device by a reading device at a specified radiofrequency. PIT tags are passive devices with a microchip that can be activated by a source emitting a voltage and generating a magnetic field. Once activated, the tag emits a signal that is then recovered by the energising source. Usually, the radiofrequencies used are low, in the range 125–135 kHz. This frequency range is a compromise between two observations: (i) as frequency increases more of the electromagnetic energy is absorbed by water or body tissues and (ii) as frequency decreases the size of the antenna needed to energise the PIT tag increases.

PIT tags are passive transponders made of an electronic component encapsulated in a glass casing. PIT tags come in various sizes, but the current optimal size for the tracking of gravel is 23.1 by 3.85 mm. This limits the size of the smallest particles that can be tagged to about 40 mm. Smaller PIT tags are under development and may become available in the near future. Tags can be activated through water or the substrate of the river bed. An antenna is used for both to supply energy to the PIT tag and to recover the signal from the tag. An antenna is basically a loop of wires that is energized by a power supply at a given frequency. The antenna can be stationary or portable depending on the objectives of the study. For a study that aims at quantifying a sediment budget in a river or at tracking the passage of clasts at a given cross-section of a stream, stationary or fixed antennas would be preferable as they register the passage of all tagged particles within the electromagnetic field generated by the antenna. Fixed antennas can be buried completely into the river bed or be partly in the bed and partly in the air. Because the antenna must be large enough to encompass the entire channel width, this application is better suited for small streams and for installations at bridges or culverts.

If the purpose of the study is to follow particles as bedload and to document the distances of movement and the location of departure and deposition over a long reach, a portable antenna that can scan the river bed is required. Because it is usually hand-held, a portable antenna is limited in size and a kit containing a battery pack, the electronics and a computer for recording the data must be carried by the operator of the antenna. The maximum power of the signal is a function of the battery that supplies the energy to the coil of the antenna. An array of antennas can be deployed as a mat into a river bed to detect the passage and location of tagged particles or fish in a river reach. The scanning of the antennas can rely on a multiplex system where each antenna is scanned in sequence.

An important advantage of the PIT tags is that they can be encrypted with a unique code that is used to identify individuals. PIT tags come in various forms: as a read-only device where the encrypted code is done by the manufacturer, as a write-once-read-many device where the user can access the device to encrypt the code once and then read it many times, and as a read–write device. In geomorphological applications, the second type of device (technically known as RI-TRP-WRHP) is often used in order to give the tracked particle a unique code or signature that can be recovered over and over again during the lifetime of the PIT tag, which can be as long as 75 years. PIT tags are highly suitable for long-term studies of sediment transport in gravel bed rivers.

The detection or read range of the system is a function of the size and power of the antenna, of the size and characteristics (e.g. the technology used to build the microchip) of the PIT tag, of the modulation of the PIT signal and of the duplex system used to recover the tag signal. Other factors that are more contingent are also important: the orientation of the tag with respect to the antenna, the presence of competing sources of electromagnetic fields (e.g. interference with other electronic devices) and the contact with ferrous material (e.g. scour chain). Lamarre *et al.* (2005) have shown that the detection range is 30 cm from the edge of a circled antenna of 0.5 m diameter powered with a 12 V battery. They also tested the depth of detection within a gravelly bed and obtained a depth of 25 cm from the bed surface. These detection ranges may vary with the size and shape of the antenna and with the power supply.

The insertion of the PIT tags into the clasts can rely on several techniques. The most secure approach is to drill a hole in the clast into which the tag is put and sealed with silicone. This protects the PIT tag very well from contacts that could break the encapsulated glass. Drilling, however, requires proper equipment that can be deployed in the field if the particles cannot be brought to the laboratory. A second method is to attach or glue the PIT tag to the clasts using silicone or cement. The tags can be inserted into a crack of the stone or into a depression on the surface of the pebble in order to protect the tag. This method is simpler to implement but it is more susceptible to breakages of the encapsulating glass of the PIT tags. Caution must be taken to select an appropriate material to attach the tag to the rock surface that is not harmful to the encapsulating glass and to the electronic device.

The advantages of PIT tags are their durability over the long term, the potential to track individual clasts marked with a unique code and the ability to recover marked stones that are exposed on the bed surface or buried in the active layer up to a depth of 25 cm. These advantages are summarized in Table 14.1. Recent experiments to track bedload particles in rivers showed that recovery rates can be as high as 94%. However, recovery rates are lower in larger rivers, as shown by Rollet *et al.* (2008), who obtained a recovery rate of only 36% in the Ain (France) where they had seeded 400 particles on a bar. PIT tags can also be applied to a wide range of particle transport problems in rivers from gravel to large woody debris. Although the use of passive transponders offers high potential for the study of gravel movement in rivers, further work is needed to optimize the application of the technique to the wide range of conditions found in rivers. In particular, the detection range could be improved through better antenna designs in order to find particles that are buried deeper than 25–30 cm. It would also be critical to estimate the burial depth of tagged particles by refining the search using two antennas. The question of the minimal size of particles that could be tagged is becoming less crucial as smaller PIT tags are becoming available. Even if the problem could partly be alleviated through the design of larger portable antennas allowing for a more efficient scan of the river bed, the application of the technique to large rivers is still a challenge.

The PIT tags method was used in to study channel dynamics, bed stability and sediment transport in Moras Creek, a 6 m wide gravel bed river located in Eastern Québec, Canada (MacVicar and Roy 2007a,b, 2011). The recovery rate exceeded 83% for four surveys (94% for the final survey) and was 66% for one survey for which anchor ice precluded a full coverage of the studied reach. The high recovery rates demonstrate that the technique can monitor the displacement of particles in a small river over a time-scale of more than 1 year, even when the bed was intensely mobilized by large morphogenic floods. The path lengths of the mobilized particles followed an exponential distribution, consistent with the hypothesis of Pyrce and Ashmore (2003) that such distributions are characteristic of small, steep channels with poorly developed pool–bar morphology. An important result concerns the relation between particle mobility and bed shear stress estimated from two different approaches: from the mean streamwise velocity and from the turbulent kinetic energy. The superposition of the mobile and immobile particles on a map of the distribution of the bed shear stress clearly shows that the most active areas of the bed correspond to where the shear stress is higher, especially when it is estimated from the mean velocity. MacVicar and Roy (2011) also showed that the mobilized particles closest to D_{50} in size are more associated with the shear stress values derived from the turbulent kinetic energy, whereas the largest particles (D_{84}) are better associated with the shear stress estimated from the average velocity.

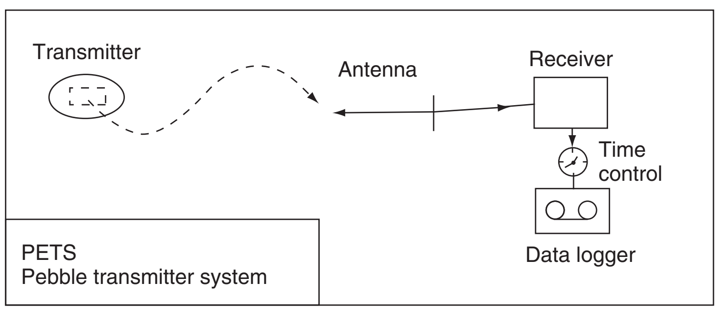
Radio tracking

This method permits the active detection of tracers during flow events. Within a tracer set, each stone emits at slightly different frequencies so that stones can be separated from one another. Few systems of this kind are available: the Lainbach (Ergenzinger *et al.* 1989) system (described below), the Toklat system (Chacho *et al.* 1989, 1994; Emmett *et al.* 1990), the Waimakariri system (Habersack 2001) and the Reynolds system (McNamara and Borden 2004; McNamara *et al.* 2008). The Toklat radio tracking system consisted of a radio transmitter, which included an antenna and battery, a radio receiver and a directional antenna. The transmitters were 18 × 72 mm in size and lasted about 10 months. For the Toklat River, nine tracers were equipped with radios transmitting at different frequencies. Also, two transmitters were equipped with a motion sensor that emitted a signal to indicate whether the particle was at rest or in motion.

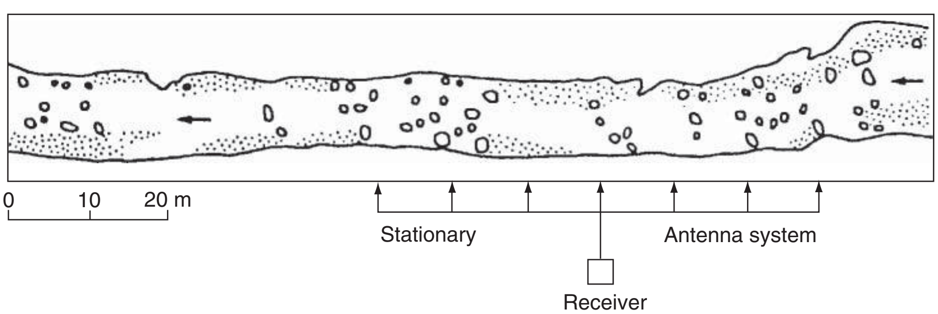
The Lainbach system (e.g. Ergenzinger *et al.* 1989; Ergenzinger and Schmidt 1990; Schmidt and Ergenzinger 1990) consisted of a transmitter, an antenna, a receiver and a data logger to store data. The transmitter emitted 2 m longwave signals, operated at a frequency of 150 MHz and could be received on the riverbank (Fig. 14.4a and b). A waterproof capsule containing a transmitter, lithium battery, antenna and mercury switch was inserted inside a hole drilled into the centre of a pebble. The capsule was 65 mm long and 20 mm in diameter and the life expectancy of the battery was about 3 months. A plug over the hole allowed battery replacement. The function of the antenna and the mercury switch was to change the emitted signals as the particle rotated.

The tagged particle also can be located during and after a flow event using antennas. There are three types of antenna: stationary, mobile and search antennas. The stationary antennas, 2 m long, are located on the riverbank (Fig. 14.4). These antennas, 5 m apart, are used to follow the passage of the tagged particles. The mobile antenna, mounted on a tripod, is carried along the study reach and used to maintain continuous contact with the particle. This antenna allows one to locate a particle to within 1 m from a distance as far as 100 m from the tracer. Finally, after flow, a special search antenna can be used to locate the exact particle position. The Lainbach study showed, as described by Einstein (1937), that the particles' displacement included phases of movement (single step) and phases of rest (rest period) (Fig. 14.5a). The observed step duration and length were very short and varied with bed characteristics and flow conditions. They also found that the distributions of both the step lengths and the rest periods (Fig. 14.5b) followed a gamma function (Ergenzinger and Schmidt 1990; Schmidt and Ergenzinger 1990, 1992; Busskamp 1993, 1994). Schmidt and Ergenzinger (1992) reported that the exponential function yielded better results than the gamma function; however, the exponential is a special case of the gamma function.

Similarly, the Waimakariri systems consist of transmitter, antennas, cables, switching box, receiver, steering computer, notebook and power supply (Habersack 2001). The battery-powered radio transmitters are inserted into gravel particles. As a function of distance, the position of the tagged particle was recorded using the intensity of the signal. For the Waimakariri River, 16 tracers were used to study step length,



(a)



(b)

Figure 14.4 (a) Sketch of the pebble radio system and (b) the antenna system along the Lainbach study site. Source: Ergenzinger *et al*, 1989. Reproduced with permission of Elsevier.

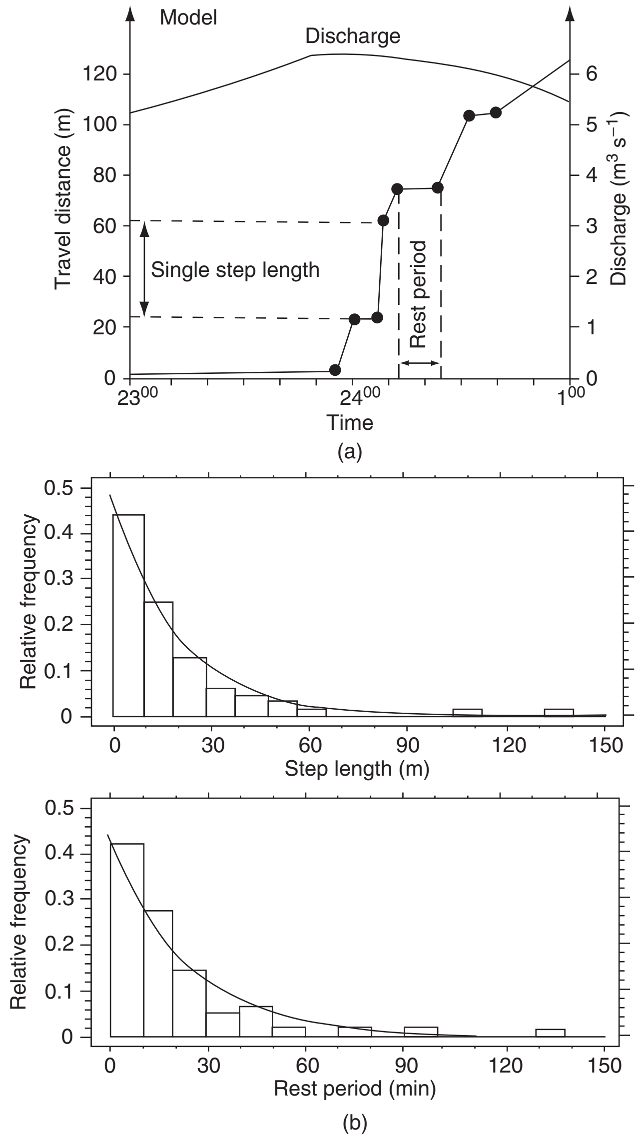


Figure 14.5 (a) Step length and rest period measurements in the Lainbach. Source: Busskamp, 1994. Reproduced with permission of Springer. (b) Distributions of step lengths and rest periods in the Lainbach. Source: Schmidt and Ergenzinger, 1992. Reproduced with permission of Wiley.

rest period and particle travel distance in a large braided river system. In principle, the Reynolds system (McNamara and Borden 2004) is similar to the other developed radio tracking systems. However, McNamara *et al.* (2008) reported a low recovery rate (26%), which was attributed to power supply problems and to the relatively large amount of sediment mobilized by an extreme event.

14.3 Conclusion

In this chapter, we have reviewed several techniques that have been used in tracing bed material load in fluvial systems. Selecting a suitable technique depends on its performance in the field, determined mainly by the recovery rate and depth of detection relative to the thickness of the active layer. Generally, methods that are reasonably successful for large material are unsuited for fine material and vice versa.

Visual detection of exotic and painted particles is probably the simplest, cheapest and most easily applied method of tracing sediment. However, the recovery rate is relatively low and is around 50% in the case of the first sediment transport event. Given that the method is limited to the bed surface, its representation of sediment transport through the scouring layer is poor. Overall, it may be concluded that these methods should be used only to obtain rough and qualitative estimates of sediment transport, but can serve very well to document the flow at which bed mobilization is initiated. For a more precise knowledge of sediment transport, it is advisable to use other methods. Although a radioactive technique is a reliable and powerful tracing technique, its toxicity makes it impractical in most circumstances. An alternative, for sand, is the fluorescent tracing method.

Magnetic tracing techniques were developed as an alternative to the paint and exotic methods for tagging coarse sediment. Artificial magnetic particles are the most widely used and are well suited to fluvial systems with discrete events during the rainy season, such that data can be obtained between events. PIT tags offer advantages over magnetic particles because they are uniquely encoded, easy to track on the river bed and will likely be improved in the future such that tagged particles can be detected even when buried in the substrate at depths greater than 25 cm.

Radio tracers are a very attractive alternative to both paint and magnetic tracing techniques. Unfortunately, high cost limits the number of tracers and hence the information obtained is limited, although very valuable. To represent the movement of natural material in a fluvial system, this method should be used in connection with other tracing techniques such as the magnetic approach.

Acknowledgements

This paper benefited from valuable comments and suggestions made by G. Mathias Kondolf, Hervé Piégay and Peter Ergenzinger in the first edition.

References

Allan, J.C., Hart, R. and Tranquili, J.V. 2006. The use of Passive Integrated Transponder (PIT) tags to trace cobble transport in a mixed sand-and-gravel beach on the high-energy Oregon coast, USA. *Marine Geology* 232: 63–86.

Ariman, J.J., Svasek, J.N. and Verkerk, B. 1960. The use of radioactive isotopes for the study of littoral drift. *Philips Technical Review* 21: 157–166.

Arkell, P.L, Leeks, G., Newson, M. and Oldfield, F. 1983. Trapping and tracing: some recent observations on supply and transport of coarse sediment from Wales. In: Collinson, J.D. and Lewin, J., eds., Modern and Ancient Fluvial Systems, *IAS Special Publication 6*, Oxford: Blackwell, pp. 107–119.

Ashworth, P.J. 1987. *Bedload Transport and Channel Changes in Gravel Bed Rivers*, PhD Thesis, University of Stirling, 352pp.

Ashworth, P.J. and Ferguson, R.I. 1989. Size selective entrainment of bedload in gravel bed streams. *Water Resources Research* 25: 627–634.

Bertoni, D., Sarti, G., Benelli, G., Pozzebon, A. and Raguseo, G. 2010. Radio Frequency Identification (RFID) technology applied to the definition of underwater and subaerial coarse sediment movement. *Sedimentary Geology* 228: 140–150.

Bradley, D.N. and Tucker, G.E. 2012. Measuring gravel transport and dispersion in a mountain river using passive radio tracers. *Earth Surface Processes and Landforms* 37: 1034–1045.

Bunte, K. and Ergenzinger, P. 1989. New tracer techniques for particles in gravel bed rivers. *Bulletin de la Société Géographique de Liège* 25: 85–90.

Bunte, K., Custer, S., Ergenzinger, P. and Spieker, R. 1987. Messung des Grobgeschiebetransportes mit der Magnettracertechnik. *Deutsche Gewässerkundliche Mitteilungen* 31: 60–67.

Busskamp, R. 1993. *Erosion, Einzellaufwege und Ruhephasen: Analysen und Modellierungen der Stochastischen Parameter des Grobgeschiebetransportes*, PhD Thesis, Free University of Berlin, 137pp.

Busskamp, R. 1994. The influence of channel steps on coarse bedload transport in mountain torrents: case study using the radio tracer technique 'PETSY'. In: Ergenzinger, P. and Schmidt, K.-H., eds., *Dynamics and Geomorphology of Mountain Rivers*, Lecture Notes in Earth Sciences 52, Berlin: Springer, pp. 129–140.

Butler, R.P. 1977. Movement of cobbles in a gravel bed stream during a flood season. *Geological Society of America Bulletin* 88: 1072–1074.

Caillot, A. 1970. Les méthodes de marquage des sédiments par des indicateurs radioactifs. *La Houille Blanche* 7: 661–674.

Caillot, A. 1983. *Guidebook on Nuclear Techniques in Hydrology*, IAEA Technical Report Series No. 91, Vienna: International Atomic Energy Agency.

Carré, D.M., Biron, P.M. and Gaskin, S.J. 2007. Flow dynamics and bedload sediment transport around paired deflectors for fish habitat enhancement: a field study in the Nicolet River. *Canadian Journal of Civil Engineering* 34: 761–769.

Chacho, E.F. Jr,, Edward, F., Burrows, R.L. and Emmett, W.W. 1989. Detection of coarse sediment movement using radio transmitters. In: *Proceedings of the XXIII IAHR Congress on Hydraulics and the Environment, Ottawa*, International Association for Hydraulic Research, pp. B-367–B-373.

Chacho, E.F. Jr, Emmett, W.W. and Burrows, R.L. 1994. Monitoring gravel movement using radiotransmitters. In: *Hydraulic Engineering '94: Proceedings of the 1994 Conference Held in Buffalo, New York, August 1–5, 1994*, New York: American Society of Civil Engineers, pp. 785–789.

Church, M. 1972. Baffin Island Sandurs: a Study of Arctic Fluvial Processes, *Geological Survey of Canada Bulletin 216*, Ottawa: Department of Energy, Mines and Resources, 208pp.

Church, M. and Hassan, M.A. 1992. Size and distance of travel of unconstrained clasts on a streambed. *Water Resources Research* 28: 233–241.

Coakley, J.P. and Long, B.F.N. 1990. *Tracing the Movement of Fine Grained Sediment in Aquatic Systems: a Literature Review*, Scientific Series No. 174, Burlington, ON: Environment Canada, Inland Waters Directorate, 21pp.

Crickmore, M.J., Tazioli, O.S., Appleby, P.G. and Oldfield, F. 1990. *The Use of Nuclear Techniques in Sediment Transport and Sedimentation Problems*, Technical Documents in Hydrology, IHP-III Project 5.2, Paris: UNESCO, 170pp.

Curtiss, G.M., Osborne, P.D. and Horner-Devine, A.R. 2009. Seasonal patterns of coarse sediment transport on a mixed sand and gravel beach due to vessel wakes, wind waves and tidal currents. *Marine Geology* 259: 73–85.

DeVries, M. 1973. Applicability of fluorescent tracers. In: *Tracer Techniques in Sediment Transport*, IAEA Technical Report Series No. 145, Vienna: International Atomic Energy Agency.

DeVries, P. 2000. *Scour in Low Gradient Streams: Patterns, Processes and Implications for the Survival of Salmonid Embryos*, PhD Thesis, University of Washington, Seattle, WA, 365pp.

Eaton, B.C., Andrews, C.A.E, Giles, T.R. and Phillips, J.C. 2010. Wildfire, morphologic change and bed material transport at Fishtrap Creek, British Columbia. *Geomorphology* 118: 409–424.

Einstein, H.A. 1937. *Bedload Transport as a Probability Problem*, PhD Thesis, ETH Zurich, English translation by Sayre, W.W., 1972, in Shen, H.W., ed., *Sedimentation*, Fort Collins, CO: Colorado State University, Appendix C.

Emmett, W.W., Burrows, R.L., Edward, F. and Chacho, E.F. Jr., 1990. Coarse particle transport in a gravel-bed river. Presented at the Third International Workshop on Gravel-bed Rivers, Florence, Italy, 24–28 September, 1990.

Ergenzinger, P. and Conrady, J. 1982. A new tracer techinque for measuring bedload in natural channels. *Catena* 9: 77–80.

Ergenzinger, P. and Custer, S.G. 1983. Determination of bedload transport using naturally magnetic tracers: first experiences at Squaw Creek, Gallatin County, Montana. *Water Resources Research* 19: 187–193.

Ergenzinger, P. and Schmidt, K.-H. 1990. Stochastic elements of bedload transport in a step-pool mountain river. In: *Hydrology in Mountainous Regions 11*, IAHS Publication No. 194, Wallingford: International Association of Hydrological Sciences, pp. 39–46.

Ergenzinger, P., Schmidt, K.-H. and Busskamp, R. 1989. The pebble transmitter system (PETS): first results of a technique for studying coarse material erosion, transport and deposition. *Zeitschrift für Geomorphologie N.F.* 33: 503–508.

Ferguson, R.I. and Hoey, T.B. 2002. Long-term slowdown of river tracer pebbles: generic models and implications for interpreting short-term tracer studies. *Water Resources Research* 38(8):17-1–17-11.

Ferguson, R.I. and Wathen, S.J. 1998. Tracer-pebble movement along a concave river profile: virtual velocity in relation to grain size and shear stress. *Water Resources Research* 34: 2031–2038.

Ferguson, R.I., Hoey, T.B., Wathen, S.J. and Werritty, A. 1996. Field evidence for rapid downstream fining of river gravels through selective transport. *Geology* 24: 179–182.

Froehlich, W. 1982. *The Mechanism of Fluvial Transport and Waste Supply into the Stream Channel in a Mountainous Flysch Catchment*, Polska Akademia Nauk, Instytut Geografii i Przestrzennego Zagospodarowania, Prace Geograficzne No. 143, Wrocław: Zakład Narodowy im. Ossolińskich.

Gintz, D. 1994. *Transportdistanzen und Raumliche Verteilung von Grobgeschieben in Abhangigkeit von Geschiebeeigenscharten und Gerinnemorphologie,* PhD Thesis, Free University of Berlin, 106pp.

Gintz, D., Hassan, M.A. and Schmidt, K.-H. 1996. Frequency and magnitude of bedload transport in a mountain river. *Earth Surface Processes and Landforms* 21: 433–445.

Gottesfeld, A.S., Hassan, M.A., Tunnicliffe, J. and Poirier, R.W. 2004. Sediment dispersion in salmon spawning streams: the influence of floods and salmon red construction. *Journal of the American Water Resources Association* 40: 1071–1086.

Habersack, H.M. 2001. Radio-tracking gravel particles in a large braided river in New Zealand: a field test of the stochastic theory of bed load transport proposed by Einstein. *Hydrological Processes* 15: 377–391.

Haschenburger, J.K. 1996. *Scour and Fill in a Gravel-bed Channel: Observations and Stochastic Models*, PhD Thesis, University of British Columbia, Vancouver.

Haschenburger, J.K. 2011a. The rate of fluvial gravel dispersion. *Geophysical Research Letters* 3: L24403.

Haschenburger, J.K. 2011b. Vertical mixing of gravel over a long flood series. *Earth Surface Processes and Landforms* 36: 1044–1058.

Haschenburger, J.K. and Church, M. 1998. Bed material transport estimated from the virtual velocity of sediment. *Earth Surface Processes and Landforms* 23: 791–808.

Hassan, M.A. 1988. *The Movement of Bedload Particles in a Gravel Bed Stream and Its Relationship to the Transport Mechanism of the Scour Layer*, PhD Thesis, Hebrew University of Jerusalem, 203pp. (in Hebrew).

Hassan, M.A. 1990. Scour, fill and burial depth of coarse material in gravel bed streams. *Earth Surface Processes and Landforms* 15: 341–356.

Hassan, M.A. 1993. Structural controls of the mobility of coarse mate rial in gravel-bed channels. *Israel Journal of Earth Sciences* 41: 105–122.

Hassan, M.A. and Church, M. 1992. The movement of individual grains on the streambed. In: Billi, P., Hey, R.E., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel Bed Rivers*, Chichester: John Wiley & Sons, pp. 159–175.

Hassan, M.A. and Church, M. 1994. Vertical mixing of coarse particles in gravel bed rivers: a kinematic model. *Water Resources Research* 30: 1173–1185.

Hassan, M.A., Schick, A.P. and Laronne, J.B. 1984. The recovery of flood-dispersed coarse sediment particle, a three dimensional magnetic tracing method, In: Schick, A.P., ed., Channel Processes – Water, Sediment and Catchment Controls, Catena Supplement 5: 153–162.

Hassan, M.A., Church, M. and Schick, A.P. 1991. Distance of movement of coarse particles in gravel bed streams. *Water Resources Research* 27: 503–511.

Hassan, M.A., Church, M. and Ashworth, P.J. 1992. Virtual rate and mean distance of travel of individual clasts in gravel-bed channels. *Earth Surface Processes and Landforms* 17: 617–628.

Hassan, M.A., Schick, A.P. and Shaw, P.A. 1995. Movement of pebbles on a sandbed river, Botswana. In: *Application of Tracers in Arid Zone Hydrology*, IAHS Publication No. 232, Wallingford: International Association of Hydrological Sciences, pp. 437–442.

Hassan, M.A., Schick, A.P. and Shaw, P.A. 1999. The transport of gravel in an ephemeral sandbed river. *Earth Surface Processes and Landforms* 24: 623–640.

Hassan, M.A., Church, M., Rempel, J. and Enkin, R.J. 2009. Promise, performance and current limitations of a magnetic bedload movement detector. *Earth Surface Processes and Landforms* 34: 1022–1032.

Hattingh, J. and Rust, I.C. 1993. Flood transport and deposition of tracer heavy minerals in a gravel-bed meander bend channel. *Journal of Sedimentary Petrology* 63: 828–834.

Hill, K.M., DellAngelo, L. and Meerschaert, M.M. 2010. Heavy-tailed travel distance in gravel bed transport: an exploratory enquiry. *Journal of Geophysical Research* 115: F00A14.

Houbrechts, G., Levecq, Y., Vanderheyden, V. and Petit, F. 2011. Long-term bedload mobility in gravel-bed rivers using iron slag as a tracer. *Geomorphology* 126: 233–244.

Hours, R. and Jeffry, P. 1959. Application des isotopes radioactifs à l'étude des mouvements de sédiments et des galets dans les cours d'eau et en mer. *La Houille Blanche* 14: 318–347.

Hubbell, D.W. and Sayre, W.W. 1964. Sand transport studies with radioactive tracers. *Journal of Hydraulics Division, American Society of Civil Engineers* 90: 39–68.

Ingle, J.C. Jr., 1966. *The Movement of Beach Sand*, Amsterdam: Elsevier, 221pp.

Keller. E.A. 1970. Bedload movement experiments, Dry Creek, California. *Journal of Sedimentary Petrology* 40: 1339–1344.

Kennedy, V.C. and Kouba, D.L. 1970. *Fluorescent Sand as a Tracer of Fluvial Sediment*, US Geological Survey Professional Paper 562-E, Washington, DC: US Government Printing Office, pp. E1–E13.

Knighton, A.D. 1989. River adjustment to change in sediment load: the effects of tin mining on the Ringarooma River, Tasmania, 1875–1984. *Earth Surface Processes and Lamdforms* 14: 333–359.

Kondolf, M.G. and Matthews, W.V.G. 1986. Transport of tracer gravels on a coastal California river. *Journal of Hydrology* 85: 265–280.

Lamarre, H. and Roy, A.G. 2008a. A field experiment on the development of sedimentary structures in a step-pool reach. *Earth Surface Processes and Landforms* 33: 1064–1081.

Lamarre, H. and Roy, A.G. 2008b. The role of morphology on the displacement of particles in a step-pool river system. *Geomorphology* 99: 270–279.

Lamarre, H., MacVicar, B. and Roy, A.G. 2005. Using passive integrated transponder (PIT) tags to investigate sediment transport in gravel-bed rivers. *Journal of Sedimentary Research* 75: 736–741.

Langedal, M. 1997. The influence of a large anthropogenic sediment source on the fluvial geomorphology of Knabeana–Kvina Rivers, Norway. *Geomorphology* 19: 117–132.

Laronne, J.B. and Carson, M.A. 1976. Interrelationship between morphology and bed material transport for a small gravel-bed channel. *Sedimentology* 23: 67–85.

Laronne, J.B. and Duncan, M.J. 1992. Bedload transport paths and gravel bar. In: Billi, P., Hey, R.E., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel Bed Rivers*, Chichester: John Wiley & Sons, pp. 177–202.

Lean, G.H. and Crickmore, M.J. 1960. *The Laboratory Measurement of Sand Transport Using Radioactive Tracers*, Wallingford: Department of Science and Industrial Research, Hydraulic Research Station, 26pp.

Lekach, J. 1992. *Bedload Movement in a Small Mountain Watershed in an Extremely Arid Environment*, PhD Thesis, Hebrew University of Jerusalem, 131pp. (in Hebrew).

Lenzi, M.A. 2004. Displacement and transport of marked pebbles, cobbles and boulders during floods in a steep mountain stream. *Hydrological Processes* 18: 1899–1914.

Leopold, L.B. and Emmett, W.W. 1981. Some observations on the movement of cobbles on a stream bed. In: Walling, D.E., ed., *Erosion and Sediment Transport Measurements (Proceedings of the Florence Symposium, June 1981)*, IAHS Publication No. 133, Late Papers – Poster

Session, Wallingford: International Association of Hydrological Sciences, pp. 49–59.

Leopold, L.B., Emmett, W.W. and Myrick, R.M. 1966. *Channel and Hillslope Processes in a Semi-arid Area, New Mexico*, US Geological Survey Professional Paper 352G, Washington, DC: US Government Printing Office, pp. 193–253.

Lewin, J. and Macklin, M.G. 1987. Metal mining and floodplain sedimentation. In: Gardiner, V., ed., *International Geomorphology 1986: Proceedings of the First International Conference on Geomorphology, Part I*, Chichester: John Wiley & Sons, pp. 1009–1027.

Liébault, F., Bellot, H., Chapuis, M., Klotz, S. and Deschatres, M. 2012. Bedload tracing in high-sediment load mountain stream. *Earth Surface Processes and Landforms* 37: 385–399.

Macklin, M.G. and Lewin, J. 1989. Sediment transfer and transformation of an alluvial valley floor: the river South Tyne, Northumbria, UK. *Earth Surface Processes and Landforms* 14: 233–246.

Macklin, M.G., Rumsby, B.T. and Newson, M.D. 1992. Historical floods and vertical accretion of fine-grained alluvium in the Lower Tyne Valley, north England. In: Billi, P., Hey, R.E., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel Bed Rivers*, Chichester: John Wiley & Sons, pp. 573–589.

MacVicar, B.J. and Roy, A.G. 2007a. Hydrodynamics of a forced riffle pool in a gravel bed river: 1. Mean velocity and turbulence intensity. *Water Resources Research* 43: W12401.

MacVicar, B.J. and Roy, A.G. 2007b. Hydrodynamics of a forced riffle pool in a gravel bed river: 2. Scale and structure of coherent turbulent events. *Water Resources Research* 43: W12402.

MacVicar B.J. and Roy A.G. 2011. Sediment mobility in a forced riffle-pool. *Geomorphology* 125: 445–456.

McNamara, J.P. and Borden, C. 2004. Observations on the movement of coarse gravel using implanted motion-sensing radio transmitters. *Hydrological Processes* 18: 1871–1884.

McNamara, J.P., Oatley, J.A., Kane, D.L. and Hinzman, L.D. 2008. Case study of a large summer flood on the North Slope of Alaska: bedload transport. *Hydrology Research* 39: 299–308.

Michalik, A. and Bartnik, W. 1986. Beginning of bedload motion in rivers. In: Wang, S.Y., Shen, H.W. and Ding, L.Z., eds., *River Sedimentation, Volume 3: Proceedings of the Third International Symposium on River Sedimentation: Central Theme, Estuarine and Coastal Sedimentation*, Jackson, MS: School of Engineering, University of Mississippi, pp. 177–186.

Michalik, A. and Bartnik, W. 1994. An attempt at determination of incipient bedload motion in mountain streams. In: Ergenzinger, P. and Schmidt, K.-H., eds., *Dynamics and Geomorphology of Mountain Rivers*, Lecture Notes in Earth Sciences 52, Berlin: Springer, pp. 289–300.

Mosley, M.P. 1978. Bed material transport in the Tamaki River near Dannevirke, North Island, New Zealand. *New Zealand Journal of Science* 21: 619–626.

Nelson, D.E. and Coakley, J.P. 1974. *Techniques for Tracing Sediment Movement*, Environment Canada, Scientific Series No. 32, Burlington, ON: Inland Waters Directorate, Canada Centre for Inland Waters, 40pp.

Nichols, M.H. 2004. A radio frequency identification system for monitoring coarse sediment particle displacement. *Applied Engineering in Agriculture* 20: 783–787.

Nir, D. 1964. Les processus érosifs dans le Nahal Zine (Neguev septentrional) pendant les saisons pluvieuses. *Annales de Géographie* 73: 8–20.

Oldfield, F., Rummery, T.A., Thompson, R. and Walling, D.E. 1979. Identification of suspended sediment sources by means of magnetic measurements: some preliminary results. *Water Resources Research* 15: 211–217.

Oldfield, F., Thompson, F.R. and Dickson, D.P.E. 1981. Artificial enhancement of stream bedload: a hydrological application of superparamagnetism. *Physics of the Earth and Planetary Interiors* 26: 107–124.

Petersen, B.R. 1960. Some radioactive surface labeling methods. *Ingeniøren* 4: 99–102.

Petit, F. 1987. The relationship between shear stress and the shaping of the bed of a pebble-loaded river, La Rulles, Ardenne. *Catena* 14: 453–468.

Pyrce, R. and Ashmore, P. 2003. Bed particle path length distributions and channel morphology in gravel-bed streams. *Geomorphology* 56: 167–187.

Ramette, M. and Heuzel, M. 1962. Le Rhône à Lyon: étude de l'entrainement des galets à l'aide de traceurs radioactifs. *La Houille Blanche* 6: 389–399.

Rathbun, R.E. and Kennedy, V.C. 1978. *Transport and Dispersion of Fluorescent Tracer Particles for the Dune-bed Condition, Atrisco Feeder Canal Near Bernalillo, New Mexico*, US Geological Survey Professional Paper 1037, Washington, DC: US Government Printing Office, 95pp.

Rathbun, R.E., Kennedy, V.C. and Culbertson, J.K. 1971. *Transport and Dispersion of Fluorescent Tracer Particles for the Flat-bed Condition, Rio Grande Conveyance Channel, Near Bernardo, New Mexico*, US Geological Survey Professional Paper 562-I, Washington, DC: US Government Printing Office, 56pp.

Reid, I., Brayshaw, A.C. and Frostick, L.E. 1984. An electromagnetic device for automatic detection of bedload motion and its field applications. *Sedimentology* 31: 269–276.

Ritter, J.R. 1967. *Bed Material Movement, Middle Fork Eel River, California*, US Geological Survey Professional Paper 575, Washington, DC: US Government Printing Office, pp. C219–C221.

Rollet, A., MacVicar, B., Piégay, H. and Roy, A. 2008. A comparative study on the use of passive integrated transponders to estimate sediment transport: first results. *La Houille Blanche* 4: 110–116.

Rovira, A. and Kondolf, G.M. 2008. Bed mobility on the Deschutes River, Oregon: tracer gravel results. *Geodinamica Acta* 21: 11–22.

Roy, A.G. and Bergeron, N. 1990. Flow and particle paths at a river confluence with coarse bed material. *Geomorphology* 3: 99–112.

Rummery, T.A., Oldfield, F., Thompson, R. and Newson, M. 1979. Magnetic tracing of stream bedload. *Geophysical Journal of the Royal Astronomical Society* 57: 278–279.

Sayre, W.W. and Hubbell, D.W. 1965. *Transport and Dispersion of Labeled Bed Material: North Loup River, Nebraska*, US Geological Survey Professional Paper 433-C, Washington, DC: US Government Printing Office, 48pp.

Schick, A.P. and Sharon, D. 1974. *Geomorphology and Climatology of an Arid Watershed, Report*, Jerusalem: Department of Geography, Hebrew University of Jerusalem, 161pp.

Schick, A.P., Lekach, J. and Hassan, M.A. 1987. Vertical exchange of coarse bedload in desert streams. In: Frostick, L.E. and Reid, I., eds., *Desert Sediments: Ancient and Modern*, Geological Society Special Publication No. 35, London: Geological Society, pp. 7–16.

Schmidt, K.-H. and Ergenzinger, P. 1990. Radiotracer und Magnettracer – die Leistungen neuer Meßsysteme für die Fluviale Dynamik. *Die Geowissenschaften* 8: 96–102.

Schmidt, K.-H. and Ergenzinger, P. 1992. Bedload entrainment, travel lengths, step lengths, rest periods – studies with passive (iron, magnetic) and active (radio) tracer techniques. *Earth Surface Processes and Landforms* 17: 147–165.

Sear, D.A. 1992. Sediment transport processes in pool-riffle sequences in a river experiencing hydropower regulation. In: Billi, P., Hey, R.E., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel Bed Rivers*, Chichester: John Wiley & Sons, pp. 629–650.

Sear, D.A. 1996. Sediment transport processes in pool-riffle sequences. *Earth Surface Processes and Landforms* 21: 241–262.

Sear, D.A., Lee, M.W.E., Oakey, R.J., Carling, P.A. and Collins, M.B. 2000. Coarse sediment tracing technology in littoral and fluvial environments: a review. In: Foster, I.D.L., ed., *Tracers in Geomorphology*, Chichester: John Wiley & Sons, pp. 21–55.

Shteinman, B., Berman, T., Inbar, M. and Gaft, M. 1997. A modified fluorescent tracer approach for studies of sediment dynamics. *Israel Journal of Earth Sciences* 46: 107–112.

Slaymaker, H.O. 1972. Patterns of present sub-aerial erosion and landforms in mid-Wales. *Transactions of the Institute of British Geographers* 55: 47–68.

Spieker, R. and Ergenzinger, P. 1990. New developments in measuring bedload by magnetic tracer technique. In: Walling, D.E., Yair, A. and Berkowicz, S., eds., *Erosion, Transport and Deposition Processes*, IAHS Publication No. 189, Wallingford: International Association of Hydrological Sciences, pp. 169–178.

Stelczer, K. 1968. Investigation of bedload movement. In: *Current Problems in River Training and Sediment Movement, Symposium, Budapest*, 9pp.

Stelczer, K. 1981. *Bedload Transport: Theory and Practice*, Littleton, CO: Water Resources Publications, 295pp.

Stott, T. and Sawyer, A. 2000. Clast travel distances and abrasion rates in two coarse upland channels determined using magnetically tagged bedload. In: Foster, I.D.L., ed., *Tracers in Geomorphology*, Chichester, John Wiley & Sons, pp. 389–399.

Takayama, S. 1965. Bedload movement in torrential mountain streams. *Tokyo Geographical Papers* 9: 169–188 (in Japanese).

Thorne, R. and Lewin, J. 1979. Bank processes, bed material movement and planform development in a meandering river. In: Rhodes, D.D. and Williams, G.P., eds., *Adjustment of the Fluvial System (Binghampton Symposia in Geomorphology)*, Dubuque, IA: Kendall/Hunt Publishing, pp. 117–137.

Tunnicliffe, J., Gottesfeld, A.S. and Mohamed, M. 2000. High resolution measurement of bedload transport. *Hydrological Processes* 14: 2631–2643.

Walling, D.E., Peart, M., Oldfield, F. and Thompson, R. 1979. Identifying suspended sediment sources by magnetic measurements on filter paper residues. *Nature* 281: 110–113.

Warburton, J. and Demir, T. 2000. Influence of bed material shape on sediment tramsport in gravel-bed rivers: a field experiment. In: Foster, D.L., ed., *Tracers in Geomorphology*, Chichester: John Wiley & Sons, pp. 401–410.

Wathen, S.J., Hoey, T.B. and Werritty, A. 1997. Quantitative determination of the activity of within-reach sediment storage in a small gravel-bed river using transit time and response time. *Geomorphology* 20: 113–134.

Wilcock, P.R., Barta, A.F., Shea, E.E, Kondolf, G.M., Matthews, W.V.G. and Pitlick, J.C. 1996a. Observations of flow and sediment entrainment on a large gravel-bed river. *Water Resources Research* 32: 2897–2909.

Wilcock, P.R., Kondolf, G.M., Matthews, W.V.G. and Barta, A.F. 1996b. Specification of sediment maintenance flows for a large gravel-bed river. *Water Resources Research* 32: 2911–2921.

Wilson, D.W., Sbatella, G.M., Wang, Q.Q. and Miller, S.D. 2010. Suitability of passive integrated transponder (PIT) tags for tracking weed seed movement in soils. *Weed Technology* 24: 386–391.

Wong, M., Parker, G., DeVries, P., Brown, T.M. and Burges, S.J. 2007. Experiments on dispersion of tracer stones under lower-regime plane-bed equilibrium bed load transport. *Water Resources Research* 43: W03440.

Yano, K., Tsuchiya, Y. and Michiue, M. 1969. Tracer studies on the movement of sand and gravel. In: *Proceedings of the XIII Congress of the IAHR,* Kyoto, vol. 2, International Association for Hydraulic Research, pp. 121–129.

CHAPTER 15

Sediment transport

D. Murray Hicks[1] and Basil Gomez[2]
[1]NIWA, Christchurch, New Zealand
[2]University of Hawaii, Honolulu, HI, USA

15.1 Introduction

Life would be much simpler for river engineers and fluvial geomorphologists if all channels had rigid boundaries and the water remained free of suspended material. Loosen the boundaries and throw sediment into a river, however, and one is immediately faced with a host of problems, issues and questions to which there are often no exact answers. How far will this bank erode? How will scour affect that bridge pier? How much time will pass before a reservoir fills with sediment? To address these and many other questions, knowledge of the physical properties of the entrained sediment and measurements and/or calculations of sediment transport are required.

The tools available to acquire this information are many and varied; they must accommodate the question under consideration, different modes of sediment transport, the physical limitations of working in rivers during floods, when most sediment transport occurs (Nelson and Benedict 1950), and the fact that sediment transport in rivers invariably tends to show substantial spatial and temporal variability (Ashmore and Day 1988; Meade *et al.* 1990; Church *et al.* 1999). Consequently, in many situations, there may be no perfect tool available and several approaches must be applied to increase confidence in a result (Wren *et al.* 2000) or at least to set upper and lower bounds to it.

The mode of sediment transport, that is, whether the sediment is moving in a rolling or saltating mode or in suspension (Abbott and Francis 1977), is a primary discriminator. Suspended load, which is fine grained and dispersed throughout the flow field, demands a different measurement approach than does the coarser bedload, which is confined to a narrow zone immediately above the bed. The supply of sediment is also important. For example, in many rivers the concentration of suspended material tends to be limited by the supply of fine sediment to the channel rather than by the capacity of the flow to support it in suspension (e.g. Hicks *et al.* 2000). While the bedload is more commonly constrained by a river's transport capacity, transport rates may also be limited by sediment availability (e.g. Milhous 1973; Hayward 1980; Jackson and Beschta 1982; Gomez 1991; Lenzi *et al.* 1999).

The modes of transport, the processes that disperse sediment within a river and the factors that affect sediment supply all contribute to substantial spatial and temporal variation in the sediment load. Since measuring the sediment load everywhere continuously is impractical (if not impossible), this variation must be sampled. Hence an appropriate sampling strategy is a key component of any approach that attempts to measure sediment transport by direct means.

We begin by reviewing the fundamental concepts of transport mode, sediment supply, capacity and competence. Next, we focus on the tools available for determining the suspended load, the bedload and assessing the total sediment load. We then consider the case of sedimentation in reservoirs, which retain much or all of the inflowing sediment and, thus, afford a unique opportunity to measure time-averaged sediment transport. Finally, we discuss sediment monitoring programme design.

15.2 Basic concepts

First, it is appropriate to define what sediment is and what it is not. In the context of this chapter, sediment is particulate material covering the size range from about 0.5 μm to boulders. Typically, it comprises rock fragments or solitary mineral grains, but it also includes organic material (e.g. leaf fragments). However, in addition to a sediment load, streams and rivers also carry a 'dissolved' load in solution. This comprises constituents that are either truly dissolved in the stream water (as ions, sourced from, e.g., chemical weathering) or are so fine that they pass through the filters conventionally used to trap 'sediment' (the conventional size boundary between particulate and dissolved load is set at 0.45 μm). The dissolved load is typically less than the sediment load, although sometimes (e.g. from limestone terrain in tropical areas) it may dominate the total material flux from a river basin. The dissolved load is also typically much more thoroughly mixed through the flow than the sediment load, it is more amenable to being monitored with instruments and it often shows different relationships with flow (e.g. its concentration is often higher at baseflows than during runoff events).

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

Therefore, the methods and tools for quantifying dissolved load (e.g. Walling 1984) differ somewhat from those used for sediment and are not addressed here.

The capacity of a river determines the maximum concentration of sediment (i.e. mass of sediment per unit volume of water or per unit area of bed) that can be moved downstream. This is limited by the ability of the flowing water to disperse the sediment, either through turbulence or by traction (Bagnold 1966). Flow competence relates to the maximum size of sediment that can be moved by a given flow condition (Nevin 1946). Often, the supply of sediment to a river channel is less than its sediment transport capacity and the river is termed 'supply limited'. A variety of intrinsic (e.g. the cohesive strength of the bed and bank material) or extrinsic (e.g. the efficacy of sheet or other erosion processes) factors may combine to limit sediment supply (Nanson 1974; Walling 1974). Competence can also limit sediment supply, for example, where the bed has developed an armour layer (Gomez 1983).

Traditionally, the sediment load of a river has been subdivided by source or by mode of transport (Einstein *et al.* 1940; Fig. 15.1a); although for most practical purposes the method of measurement determines how fluvial sediment loads are reported. By source, the total load is split between bed-material load and washload. The bed-material load is derived from the river bed and is typically sand or gravel sized; its concentration is directly related to a river's transport capacity. The washload consists of sediment that has been flushed into the river from upland sources and is sufficiently fine grained that the river is always competent to entrain it in suspension. Consequently, only trace quantities of washload material are found in the bed material, even if the washload dominates the total load. Typically, the washload comprises clay, silt and up to fine sand grades, although in steep, headwater streams it may also be considered to include coarse sand and even pebbles trapped in the interstices between boulders (in fact, any sediment that would be suspended if exposed to the flow). Generally, washload concentration is dependent on the relative rates of supply of water and sediment to the channel. Being finer grained, it is rarely capacity limited; indeed, when the concentration of mud-rich washload becomes sufficiently large (several hundred thousand parts per million), the fluid properties change from those of a water flow to those of a hyperconcentrated or debris flow (Costa 1988).

By mode of transport, the sediment load is divided into suspended load and bedload. The suspended load is dispersed in the flow by turbulence and is carried for considerable distances without touching the bed. It is usually fine sand, silt and clay; in terms of source, it is largely derived from the washload and the finer fractions of the bed material. The bedload is typically coarser sediment moving in almost continuous contact with the

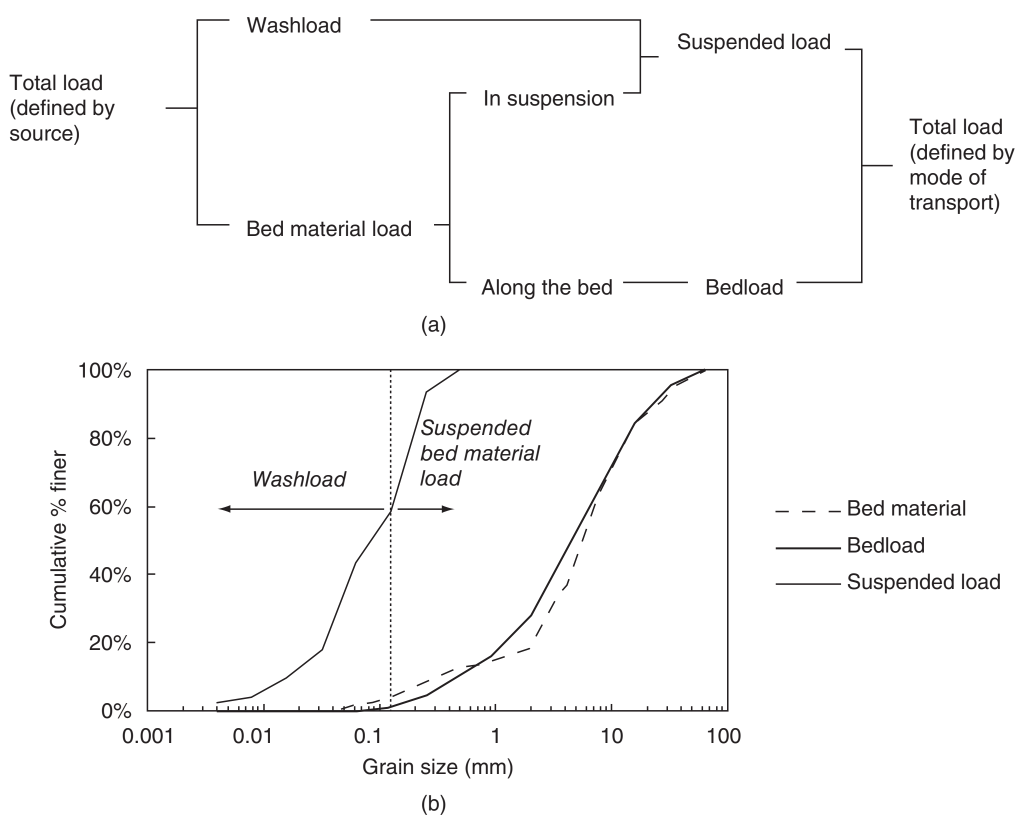


Figure 15.1 (a) Breakdown of stream sediment load in terms of sediment source and mode of transport. (b) Size grading of suspended load (average from eight samples), bedload (average of nine gaugings) and bed-material (average of nine samples) for Shotover River, New Zealand.

bed, rolling, sliding or saltating under the tractive force exerted by the water flow. In practice, particularly where sand comprises a large proportion of the total load, the boundary between bedload and suspended load blurs. Downstream through a drainage basin, the bed material generally becomes finer through the action of sorting and abrasion; in consequence, the suspended load increasingly dominates over the bedload.

To illustrate these concepts of load classification, Fig. 15.1b compares size gradings of the bed material, bedload and suspended load sampled from the Shotover River, which drains a 1000 km^2 basin in the South Island, New Zealand. Note that the bed material is bimodal, containing a dominant gravel mode that matches the gravel bedload, a significant fine–medium sand mode that matches the coarser suspended load mode and negligible quantities of sediment in the clay to very fine sand range (i.e. finer than 0.125 mm). The latter fraction, however, constitutes approximately 60% of the suspended load and may be regarded as the washload.

The differences between the supply and mode of transport of the suspended and bed loads are reflected in the different approaches employed to determine them. Because the suspended load often is related more to the sediment supply than transport capacity, it must generally be measured directly. In contrast, the bedload, which is typically controlled by the transport capacity, may (in theory, if not in practice) be more readily estimated using a theoretical or empirical approach.

15.3 Suspended load sampling and monitoring

Overview

The tools used to determine the suspended load vary with the problem to be addressed. A key control on the approach used is the time base of the problem: does the problem require continuous time-series data, event-based results or simply long-term statistics, such as the mean annual sediment yield? This is important because the effort required to conduct a single 'instantaneous' measurement of the suspended load, properly sampled across the channel, is impractical to sustain on an ongoing basis. Hence trade-offs, or simplifications, have to be made when temporal detail is required (Wren *et al.* 2000).

In this section, we first review the requirements of a single suspended sediment 'gauging' which adequately samples the cross-section spatial variability in load. We then look at strategies for collecting and analysing data on a continuous basis. Next, we consider the sediment rating and related approaches, where the interest is only in aspects of the long-term average load. We then consider sediment yields on an event basis and address methods for determining suspended load particle size. Last, we discuss synoptic sampling, where a spatial overview may be required on suspended sediment concentrations, perhaps during an extreme flood.

Suspended sediment gaugings

A suspended load gauging requires the spatially distributed measurement of both sediment concentration and water velocity. Strictly, the suspended sediment discharge, or flux (q_s), past a single vertical in a river cross-section is determined from

$$q_s = \int c_s(z)v_s(z)\mathrm{d}z \tag{15.1}$$

where c_s and v_s are the concentration and downstream velocity, respectively, of the suspended sediment. Both c_s and v_s vary with depth (z direction). The variation of c_s with depth depends on the intensity of turbulence and the fall velocity of the sediment. For a given level of turbulence, finer sediment (silt and clay), with a lower fall velocity, tends to be mixed more uniformly over the flow depth, whereas coarser sediment (sand fractions) tends to be concentrated near the bed. In practice, it is usually assumed that the sediment moves at the same velocity as the water, i.e. $v_s = v$, (where v is the streamwise fluid velocity), even though there may be a significant slip velocity in the case of sand grains (Aziz 1996). With this approximation,

$$q_s = \int c_s(z)v(z)\mathrm{d}z \tag{15.2}$$

Practically, the integral in eqn. (15.2) can be determined in two ways. The first is to collect point samples of water and to make point velocity measurements at intervals over the flow depth, plot profiles of concentration and velocity (Fig. 15.2), then integrate the product $c_s(z)v(z)$. The point samples must be collected with a properly-designed 'point sampler', such as the US P-61 (Fig. 15.3b). These cable-suspended samplers, comprising a brass bomb and an internal glass sample bottle, have a solenoid-operated valve to control the water-sample capture and are designed so that they sample isokinetically (i.e. they accept a sample at the ambient water velocity). If this was not the case, the sampled concentration would be biased for sand grade sediment. Typically, 6–8 points are required to define a point sampled profile.

The second, and simpler, way is to use a 'depth-integrating' sampler, such as the US DH48. This is similar to a point sampler except that the inlet nozzle is kept open while the sampler is traversed at a uniform rate from the water surface to the bed and back again. As it traverses the flow depth, it samples isokinetically and so performs a mechanical integration of the $c_s(z)v(z)$ product. The mass of sediment collected in the sample bottle, when multiplied by the ratio (flow depth)/(sampler nozzle area × sampling time), provides a direct estimate of q_s at the sampling vertical (in units g s^{-1} m^{-1}). However, it is more usual to determine q_s from the product of the sediment concentration in the sample bottle at the end of the traverse, termed the 'discharge-weighted mean concentration', c_q, and the unit water discharge as determined with a current meter. With the depth-integrating approach, the speed with which the sampler traverses the water column must be fast enough that the sample bottle is not filled

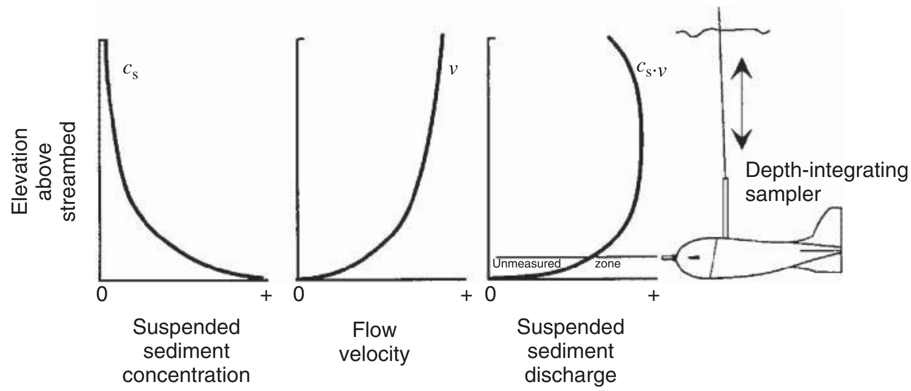


Figure 15.2 Vertical profiles of suspended sediment concentration (c_s), streamwise velocity (v) and sediment discharge (product $c_s \cdot v$). A depth-integrating sampler mechanically integrates the $c_s \cdot v$ profile if it is traversed from streambed to water surface at a uniform rate, but it misses a narrow zone next to the bed.

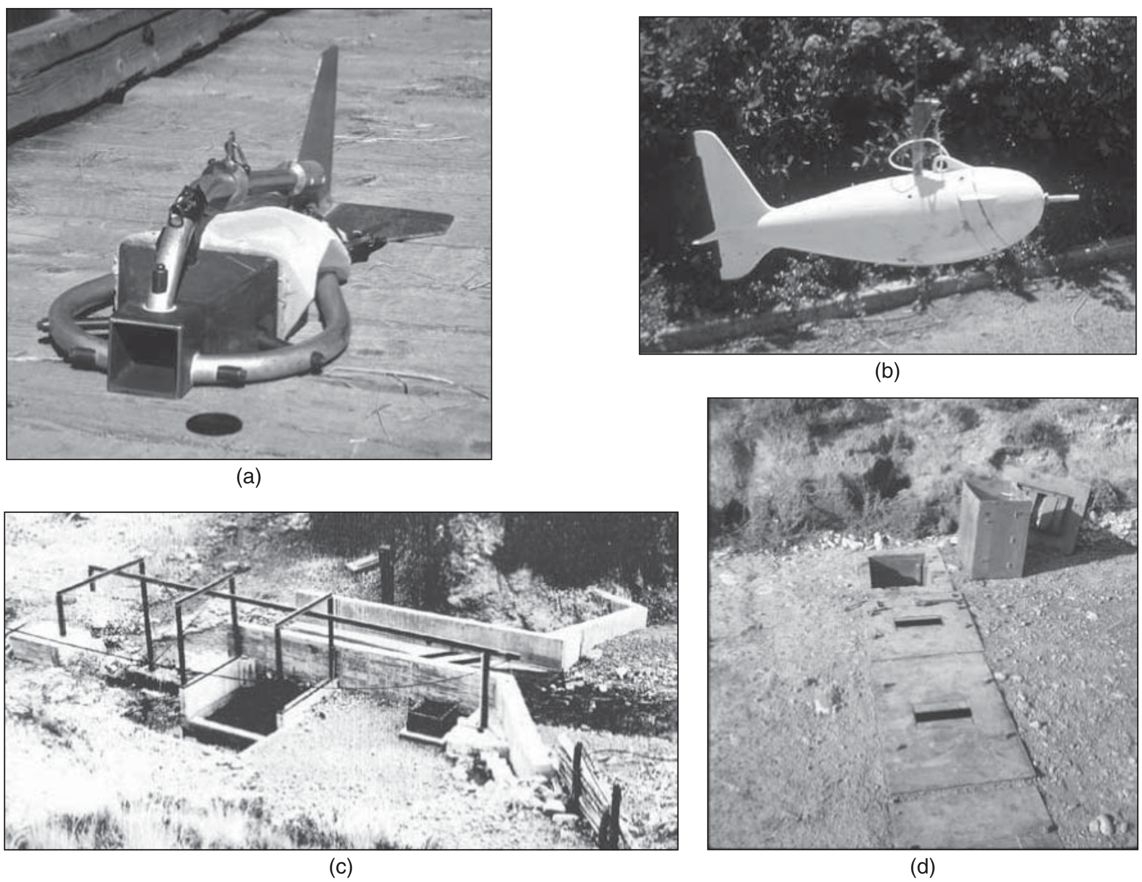


Figure 15.3 Sampling and measuring devices. (a) The Helley–Smith (1971) pressure-difference bedload sampler. (b) The US P-61 suspended sediment sampler. (c) The vortex bedload trap in Torlesse Stream (Hayward and Sutherland 1974) Source: Davies *et al*, 1974. Reproduced with permission of Elsevier. (d) The Birkbeck-type (Reid *et al.* 1980) pit trap installed in Nahel Eshtemoa (Reid *et al.* 1998); the slotted metal covers were installed to extend the lifetime of the trap during storm events. The US P-61 suspended sediment sampler, which incorporates a solenoid valve that controls nozzle operation, collects a time-integrated sample from a specific point in the channel. Like the Helley–Smith bedload sampler, it is deployed from a cable. Depth-integrated samples may be collected using US D-74 or US DH-48 samplers (not illustrated). The US DH-48 sampler is mounted on a wading rod and the Helley–Smith sampler may be deployed in a similar fashion. Vortex and Birkbeck-type bedload traps are permanent installations that may be used to continuously monitor the mass of accumulating sediment. Information about many sediment sampling and measuring devices may be obtained from the FISP Home Page: http://water.usgs.gov/fisp/.

before the traverse is complete. In some circumstances this may be too fast to sample adequately the turbulence-driven fluctuations in near-bed sand concentration (Hicks and Duncan 1997) or to avoid contorting the streamlines entering the sampler nozzle (Edwards and Glysson 1999). In such conditions, the point sampling approach, or a combination approach involving depth-integrating over limited depth spans using a point sampler, is more accurate.

Point samples may also be collected by pumping up to a surface container, and multiple lines can be used to sample simultaneously an array of points in the vertical (e.g. Van Rijn and Gaweesh 1992). With pumping, however, care is required to maintain isokinetic flow through the intake nozzle and to have velocities up the riser line substantially greater than the sediment fall velocity, otherwise a false concentration of suspended sand will be sampled. This is less of an issue with silt and clay-sized sediment.

Variations in suspended sediment load across channel may be substantial, at least for the sand fractions, which are less well mixed than the silt and clay fractions. This is dealt with by sampling at multiple verticals, preferably spaced either at equal intervals of channel width or so that sub-sections contain equal portions of the total water discharge. The total suspended sediment discharge for a section is typically found by multiplying the discharge-weighted mean concentration at each sampling vertical by the water discharge in the sub-section that each vertical represents (as obtained during an accompanying flow gauging).

Study purpose and site conditions should determine the choice of sampler type, nozzle size, sample-bottle size, traverse method and rate and the number and location of sampling verticals. Details about the standard samplers and methods developed in North America by the US Federal Interagency Sedimentation Project (FISP) can be found in technical manuals such Edwards and Glysson (1999). FISP suspended sediment samplers can be deployed using conventional flow-gauging reels from bridges or cable-cars (e.g. the US D-74) or mounted on a wading rod (US DH-48). With FISP samplers, depth-integrating samplers are designated by 'D', point samplers by 'P' and hand-held by 'H', and the number refers to the year the sampler was developed. The US P-61 is formally a point sampler, incorporating a solenoid valve that controls nozzle operation, allowing collection of a time-integrated sample from a specific point in the vertical (Fig. 15.3b). However, it can also be operated in depth-integrating mode over selected segments of the flow depth (e.g. part of the depth, one-way sampling from bed to surface or vice versa). This renders it capable of sampling deep and/or fast flows that would compromise the efficiency of standard depth-integrating samplers, since they must sample all the way between the water surface and bed and back again. Sediment sampling and measuring devices continue to evolve and up-to-date information is available on the FISP home page http://water.usgs.gov/fisp/.

Most suspended sediment samplers sample only to within 75–100 mm of the bed (Fig. 15.2). With the depth-integrating approach, the suspended sediment concentration in the unsampled zone is implicitly assumed to be equal to the mean concentration in the sampled zone. This is reasonable if the sediment is well mixed through the vertical, as silt and clay invariably are, but may underestimate the sand load since sand tends to be concentrated near the bed. Procedures for adjusting the mean concentration and size distribution of the suspended load to incorporate the unsampled zone are given in Colby and Hembree (1955) and Stevens and Yang (1989). This requires information on the concentration and size distribution of sediment in the measured zone, the flow velocity and the bed material size distribution.

Continuous monitoring

A single suspended sediment gauging, as described above, may require tens of samples and several hours to complete. Needless to say, it is usually not a measurement that can be repeated on an ongoing basis and it becomes impractical when the flow rate or sediment concentration changes rapidly. If the need is for continuous (or near-continuous) data on suspended sediment load, then the usual approach is to collect 'index' samples from a single location. Such samples may be depth-integrated from a fixed vertical or collected from a single point. Calibration measurements are required to establish a relation between the point concentration and the cross-section mean concentration.

Index samples may be collected by hand; however, in remote locations or in 'flashy' small basins, they are more often collected by an automatic pumping sampler. Auto-samplers with on-board processors, or when coupled to a programmable data logger, may be programmed to sample under various strategies, including fixed time, fixed stage change or fixed flow volume (flow-proportional) bases. The main disadvantage of auto-samplers is their relatively small number of sample bottles – typically 24–28 for the more portable samplers. This can be overcome in part by sampling on a flow-proportional basis and compositing multiple samples (up to eight) into each bottle. Other disadvantages include mechanical breakdown and limited pumping head. Typically, the pumping head capability is about 5–6 m, which constrains their application to smaller streams (although some have been modified with booster pumps, e.g. Gray and Fisk 1992). Also, auto-samplers do not sample isokinetically and so the concentrations of sand may be biased. To some extent, this bias can be removed empirically by the point to cross-section mean calibration process.

If an extended and detailed time-series record of the suspended load is required, then sensors that monitor surrogate properties of the suspended sediment concentration provide an economical option. Optical sensors are particularly attractive if the primary interest in the sediment information concerns a water clarity issue and they have been widely used to date (e.g. Walling 1977; Gippel 1989, 1995; Lawler and Brown 1992; Lewis 1996; Stott and Mount 2007). Since these do not sense sediment concentration directly, they require that a further calibration relation be established between the optical signal and the local suspended sediment concentration. The optical signal depends both on sediment concentration and on particle characteristics, notably particle size and shape. Light scattering from clay particles (< 4 μm), which are platy in shape, is 20 times more effective than from the same mass concentration of coarse silt (Foster *et al.* 1992) (Fig. 15.4a), hence optical sensors are more sensitive to washload than to suspended bed-material load. To a lesser extent, the optical signal is also sensitive to particle composition (e.g. organic particles give a different signature compared with mineral particles) and to colour-producing dissolved organic substances (Gippel 1995). As discussed later, suspended sediment particle size can vary through events and seasonally, leading to scatter in the relation between the optical signal and sediment concentration. However, at least at sites where wash load dominates the suspended load, this scatter is small compared with the range in concentration and a good calibration function is usually achieved (Fig. 15.4b). Sensor outputs typically relate linearly to sediment concentration and so linear regression can be used to derive calibration relationships. Where necessary, power relations can be fitted using linear regression of the log-transformed data, but a log-detransformation adjustment is then required (as detailed in the section 'Suspended sediment ratings').

The optical sensors may be either transmissivity (attenuance) or back-scattering (nephelometric) types. Traditionally, the back-scattering types have been better suited to monitoring sediment loads since with these the signal-to-noise ratio increases as sediment concentration increases (the reverse occurs with transmissivity sensors, which monitor light transmission over a fixed path). Even so, until recently these have been limited in range to 2000 nephelometric turbidity units (NTU), which typically corresponds to about 5000 mg L^{-1}. In the last several years, a new generation of 'smart', self-ranging sensors, of both transmissivity and back-scattering types, has become available. For the first time, these provide adequate ranges (some as high as 200, 000 mg L^{-1}) and sensitivities to monitor the full range of concentration found in most streams and rivers.

A nuisance often encountered with optical sensors is bio-fouling of the lens. This can be controlled with varying success by hand cleaning, mechanical wipers, algae-repelling polymer coats and jets that squirt chemicals or simply water onto the lens. Within limits, dual-path transmissivity sensors do not 'see' fouling because they sense the relative transmissivity over two different path lengths. Another disadvantage with optical sensors is that with only a small proportion of clay in suspension, they will not register coarser sediment – hence they are not suitable where the sand load is the primary interest.

Acoustic instruments, which monitor the intensity of sound energy back-scattered from suspended sediment, offer many of the advantages of optical sensors while being less vulnerable to bio-fouling and more robust in field deployments (Gray and Gartner 2010; Gray *et al.* 2010). They have the added advantages of a larger sampling volume and the facility to map sediment concentration along the sound beam path. Indeed, since acoustic back-scattering intensity is a by-product of acoustic Doppler current profilers (ADCPs), it is possible to use the one instrument to gauge both water and suspended sediment discharge over a cross-section (e.g. when mounted on a moving boat) or to monitor continuously both flow velocity and sediment concentration at a fixed location, such as with an up-looking, bottom-mounted ADCP (e.g. Wall *et al.* 2006).

After adjusting for various factors that influence sound beam intensity, including instrument characteristics, water properties and range-dependent sound attenuation and spreading, acoustic back-scattering (ABS) depends on both sediment concentration and particle size (Thorne *et al.* 1993; Gartner 2003; Hoitink and Hoekstra 2005; Gray and Gartner 2009). Urick (1983) considered that for particles to be detected acoustically, their circumference should exceed 0.1 times the sound wavelength, and it has been observed that ADCP instruments are more suited to detecting suspended sand rather than silt and clay (Gartner and Cheng 2001).

As with optical sensors, a limitation of acoustic sensors is distinguishing between variations in ABS due to concentration and particle size effects. This confusion can be managed by using multi-frequency instrumentation (Topping *et al.* 2007). With the more common single-frequency sensors, it is necessary to calibrate the ABS signal to the dominant sediment

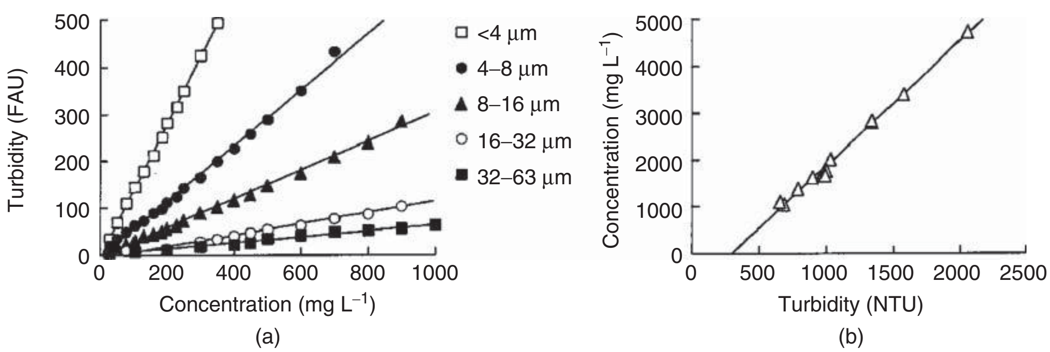


Figure 15.4 (a) Relations between turbidity (as measured by a Partech S100 dual-path sensor) and stream suspended sediment concentration for five size fractions. After Foster *et al.* (1992). (b): Relation between turbidity (as measured by an OBS-3 back-scattering sensor) and suspended sediment concentration for Waipaoa River, New Zealand. The Waipaoa's suspended load is dominantly silt and clay.

grade by collecting concurrent water samples; also, it is better to match the sensor frequency to the dominant sediment size. An alternative approach using a single-frequency sensor takes advantage of the fact that sand tends to dominate acoustic back-scattering while silt-clay dominates acoustic attenuation, hence the concentrations of both fractions can be extracted after appropriate data processing (Topping *et al.* 2007; Gray and Gartner 2010). Green *et al.* (1999) deployed both optical and acoustic sensors to target silt and sand modes, respectively.

The expectation over the next decade is for in situ ABS profiling instrumentation to emerge as the preferred surrogate for monitoring suspended sediment (Gray and Gartner 2010).

Suspended sediment ratings

If the main interest lies in determining the long-term average suspended sediment yield, then the 'sediment rating' approach offers considerable economies of sampling effort and obviates the need for continuous records of sediment concentration. The secret to its successful implementation, however, lies in a well-designed and well-implemented sampling strategy.

A sediment rating aims to represent the suspended sediment concentration as a continuous function of water discharge. There are two main approaches. The first recognizes that there is no unique relation between suspended sediment concentration, C, and water discharge, Q, and so aims to model the conditional mean concentration (as a function of water discharge) over the time period of interest. The conditional mean relation is estimated by sampling a series of concurrent measurements of water discharge and discharge-weighted sediment concentration. This relation is then combined with the water discharge record $Q(t)$ for the same period in order determine the sediment yield. In terms of accuracy, it matters little if the full flow time series is used or if it is compressed into a flow-duration table, provided that in the flow-duration table the flow range is divided into small intervals or at least the high flow range is well detailed (Miller 1951; see also Walling 1977). The greatest sources of error arise from the method used to model the relation and from the sampling strategy. The second approach involves attempting to model explicitly the suspended sediment concentration with an empirically derived multivariate relation, relating sediment concentration not only to water discharge but also to other controls or processes affecting the sediment supply, such as season, long-term trend, hysteresis of sediment delivery during storms, and so on. With this approach, time-series information is required on all of the independent variables in order to generate a long-term average sediment yield. The first approach is more common and we focus on it here.

Modelling the C–Q relation

The traditional approach to deriving a rating model (or curve) has been to plot concurrent measurements of C against Q on log–log graphs. There are several good reasons for this:

1. Log–log plots accommodate the large ranges of discharge and sediment concentration in rivers.
2. The data scatter tends to be homoscedastic (i.e. independent of discharge).
3. The underlying relation typically shows a simple power form $C = aQ^b$ (where a and b are empirical coefficients), which is linear on a log–log plot. At first, such rating equations tended to be fitted by eye, but with the arrival of personal computers and statistical analysis packages, linear regression of the log-transformed data became widely used. Ferguson (1986, 1987) pointed out that by using logarithmic data, the linear regression procedure modelled the geometric conditional mean, rather than the desired arithmetic conditional mean, hence he proposed correcting the coefficient a by the factor $\exp(s^2)$ (where s is the standard error of the estimate in natural log units), which is based on the assumption that the data scatter about the modelled line is log-normally distributed. Cohn et al. (1989) showed that if the residuals distribution was not log-normal, then Ferguson's bias-correction factor could be substantially in error. They provided an alternative method of correcting for log–log bias based on a maximum likelihood estimator. Duan's (1983) empirical 'smearing' estimator is also used for the same purpose. Crawford (1991) showed that both of the latter two correction factors improved the accuracy of the log-linear least-squares approach.

Even with appropriate bias correction, however, independent assessments of sediment yield have shown that sediment-rating assessed yields can still be in error by factors as large as 10 (Walling and Webb 1988). Such poor results can arise because the simple power law model, although appearing to fit the overall dataset reasonably well, gives a poor fit to the high discharge end of the relation (which may only be a short tail of sparse data on the right-hand end of the log–log plot, but transports the bulk of the long-term load). In such cases, other curve fitting techniques such as non-linear regression (Singh *et al.* 1986) or LOWESS (Cleveland 1979; Hicks *et al.* 2000) perform better. Essentially, LOWESS (Locally Weighted Scatterplot Smoothing) constructs a 'running' linear regression fit to the data, using a limited window (or band) of discharge and weighting each data point in inverse relation to its distance from the window centre. Figure 15.5 plots sediment rating data for the Shotover River, South Island, New Zealand, where a continuous concentration record was generated from a turbidity record by relating turbidity to concentration. Figure 15.5(a) shows the full C–Q bivariate distribution over a 6-month period, the conditional-mean concentration trend (over 50 flow bands) and a simple regression model of the log-transformed data. Note (i) how the regression model, weighted to the more numerous data at lower flows, misses the high flow tail of the plot, and (ii) the erratic form of the conditional-mean trend at higher flows, as the number of sample points per flow band decreases. Compared with the true yield (733,000 t, estimated from the full time-series of concentration and discharge records), the yield estimated by the simple regression model was 693,000 t. This increased to 716,000 t when the log bias-correction factor of Ferguson (1987)

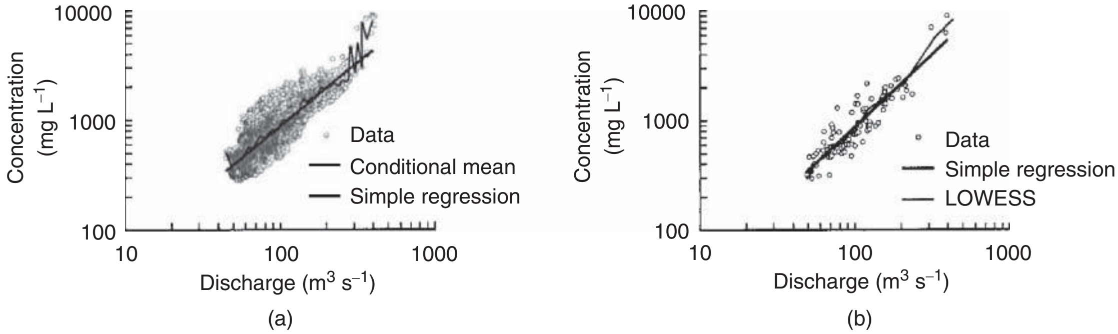


Figure 15.5 (a): Relation between suspended sediment concentration and water discharge for 6 months of hourly data from Shotover River, New Zealand, with rating relations modelled by conditional-mean concentration and simple regression. (b): Relation for a stratified random sample of 100 points, with ratings fitted to these points using simple regression and LOWESS.

was applied. Figure 15.5(b) shows a stratified random sample of 100 points (50 at flows above 100 m^3 s^{-1} and 50 at lower flows), designed to simulate a series of gaugings, plus ratings fitted to these points using simple log–log regression and LOWESS. The yields estimated by these models were 748,000 t (after log bias correction) and 743,000 t, respectively.

More recently, fuzzy logic (FL) and artificial neural network (ANN) approaches have been used to capture empirically the often complex form of the *C*–*Q* relation. The FL approach relates *C* to *Q* using multiple 'fuzzy membership' functions that partially overlap across the *Q* range (e.g. Kisi *et al.* 2006; Mianaei and Keshavarzi 2010). These functions are optimized (usually with the aid of sophisticated software 'toolboxes', e.g. the MATLAB library) to minimize the error between observed and predicted values, which is essentially what the LOWESS approach does in a smooth transition across the *Q* range. The ANN approach (e.g. Kisi 2005; Alp and Cigizoglu 2007; Mount and Abrahart 2011) develops a non-linear 'black box' predictive model that is 'trained' off an existing dataset (and, indeed, can be retrained on-the-fly with new data). Like the FL approach, the algorithms are sophisticated but accessible from software libraries and toolboxes and *C* can be related to additional hydrological variables besides *Q* (e.g. rainfall).

Sampling strategies

An often major source of error in rating relations is the data themselves. One assumption with the rating method is that there is no bias in the data collection – for example, that, for a given discharge band, there is no preference to collect samples when the concentration is less than the long-term mean concentration for those discharges. Particularly in smaller streams, there is a tendency for concentrations to be lower on flood recessions, due to an initial 'flush' of readily available sediment from the channel (e.g. Christian and Thompson 1978; Walling and Webb 1988). If the rising and falling stages of these streams are not sampled in proportion to their relative frequencies of occurrence, e.g. because the remoteness of a site prevents field parties from arriving until after the flood peak, then the sampling, and the resultant sediment rating, will be biased to the lower concentrations. If this bias can be identified, then it may be removed by developing a multivariate rating or more simply by splitting the data and deriving separate ratings (e.g. for rising and falling stage or for separate seasons). Note that in some cases, there may be a trend or non-stationary signal in the sediment supply, such as may follow from catchment land-use change or as a catchment recovers from an extreme event such as a storm or earthquake that induces catastrophic erosion (e.g. Kelsey 1980). In these cases there may be no stable, long-term sediment rating.

In practice, many existing suspended sediment datasets have been compiled with no plan or strategy that relates directly to sediment sampling considerations and unwitting bias may have been introduced. If at all possible, it is better to avoid sampling bias by designing an objective sampling strategy (Olive and Rieger 1988). Unbiased strategies include regular time-interval, random time-interval, flow-weighted probability or load-weighted probability. Simulations have shown (e.g. Thomas 1988; Walling and Webb 1988) that regular/random time-based sampling can provide poor results, since the chances of intercepting flood flows are slim. Flow- and load-weighted sampling work better, particularly where simple regression is used to derive the rating model, because they tend to result in reasonably uniform densities of data points over the flow range (e.g. at high discharges, the greater frequency of sampling tends to balance the lower frequency of occurrence). Load-weighted sampling ensures data points in the 'most effective' discharge range, i.e. that which transports the bulk of the long-term load. Thomas's (1985) selection-at-list-time (SALT) sampling method can be used to schedule sediment rating data. With this, a 'first estimate' rating is used to estimate the sediment load from the current flow and the probability of collecting a sample in a given time interval is then assigned in proportion to this estimated load. The 'estimator' rating can be tuned as data are collected. Automatic samplers, controlled by programmable data loggers coupled to a stage recorder, permit such flexible sampling strategies to be realized. In many situations, this benefit outweighs the disadvantage of having to conduct manual sampling to establish a relation between the point concentration at the sampler intake and the cross-section mean. Therefore, auto-samplers should be seriously considered where the sediment rating method is to be applied.

Where the primary purpose of auto-sampling is to calibrate surrogate records (e.g. turbidity, ABS) to sediment concentration, the auto-samples can also be used to develop a Q–C rating, thus enabling an auxiliary estimate of the suspended load. This can be used to verify the primary, surrogate-based load estimate and may also be used to patch gaps in the surrogate record.

A more direct, but less commonly used, alternative to the sediment rating method for estimating the time-averaged suspended load is the 'direct estimation' approach (Cohn 1995). This involves summing a series of selectively sampled but unbiased load estimates that are derived with an auto-sampler controlled by a programmable data logger. Examples include the time-stratified (Thomas and Lewis 1993a), flow-stratified (Thomas and Lewis 1995) and SALT (Thomas 1985) approaches.

Event suspended sediment yields

In certain situations, there is greater interest in suspended sediment yields from individual runoff events than in the long-term average yield. One example is the filling of sediment retention dams, which are frequently placed in basins undergoing urban development or timber harvesting in order to limit sediment exports. In such situations, knowledge of the magnitude–frequency distribution of the event sediment yields is important for designing sediment trap capacities and for setting limits on sediment releases to downstream waterways. Event-yield magnitude–frequency relations are also useful discriminators of land-use effects on sediment yields.

Determining a magnitude–frequency relation for event sediment yields at a site ideally requires continuously monitoring stream sediment loads (or reservoir deposition – see later) for a period of years. With a time series of stream loads, discrete sediment-yield events can be totalled on the basis of discrete quickflow events. The analysis then proceeds in a similar fashion to undertaking a peaks-over-threshold (partial duration) analysis for event peak flows (e.g. Haan 1977): the event yields above a threshold size are ranked, assigned a return period (T) using a 'plotting equation', e.g. $T = m/n$ (where m is the event rank and n the number of years of record), then modelled with an appropriate distribution.

Where no long record of continuous sediment load data is available, an alternative approach is to establish a relation between event sediment yield and some correlated index of the event magnitude that is more easily monitored. Event peak flow typically correlates well with event sediment yield and can be used for this purpose (e.g. Neff 1967; Hicks 1994) (Fig. 15.6a).

As with 'instantaneous' sediment ratings, care is required in modelling the event yield versus peak flow relation and in extrapolating it outside the range of the data. Unrealistic extrapolation can induce large errors in the yields estimated for extreme events. Parker and Troutman (1989) outlined an approach that incorporates the uncertainty in the event yield versus peak flow relation. They used a log-Pearson type III distribution to model the probability density function of the annual peak flows (Y) and a quadratic regression relation between the logarithms of annual flood peaks and associated sediment yields (Y_s). They assumed a normal distribution of the errors in the Y versus Y_s regression relation in order to estimate the conditional probability density of the event sediment yield given a peak value. Finally, they combined the functions for the probability of Y and the conditional probability of Y_s given Y to derive a function which, when integrated numerically, predicted the sediment yield for a given return period. Although Parker and Troutman (1989) dealt with annual maxima events, their approach could potentially also be applied to a peaks-over-threshold series.

An event sediment–rating relation can be compiled over 1 or 2 years, allowing for a good range of events to be sampled and for the relation to remain 'stationary' over the period of flow record used for the return period analysis (i.e. the sediment sources and erosion processes in the basin do not change appreciably, such as might occur during a land-use conversion). Automatic pumping samplers or turbidity sensors are well suited for such short-term event-sampling deployments. As already discussed, use of auto-samplers or turbidity sensors requires a phase of manual sampling to calibrate point measurements to cross-section mean values. Lewis (1996) compared strategies for collecting auto-samples for calibrating turbidity versus concentration relations during runoff events.

The effect of land use on event sediment yield magnitude–frequency relations was demonstrated by Hicks (1994) using storm yield data from four small basins around Auckland, New Zealand. The overall position of a data series on the magnitude–frequency plot (Fig. 15.6b), reflecting the sediment yield per unit area during an event of given return period, indicates the overall availability of sediment in the basin. This was higher by an order of magnitude in the basin undergoing urban development, owing to the ready sediment supply from earthworks and road construction. The steepness of the data series reflects the continuity of the sediment supply during larger events. The plot was steepest for the urbanizing basin, where sediment was abundant, flattest for the mature urban basin, where sediment became exhausted during large events, and had intermediate slopes for pasture and market-gardening basins. The mature urban basin yielded more sediment than the pasture basin during sub-monthly events, but the reverse was true with less frequent events. Such a plot is more informative than a simple comparison of average annual yields.

Event-based sediment yields may also be obtained from analysis of reservoir strata, as discussed in Section 15.6.

Suspended sediment particle size

Particle size information on suspended sediment is important for two main reasons. First, it is a primary control on entrainment and deposition and so affects the degree of mixing within a river cross-section, downstream sorting and settling

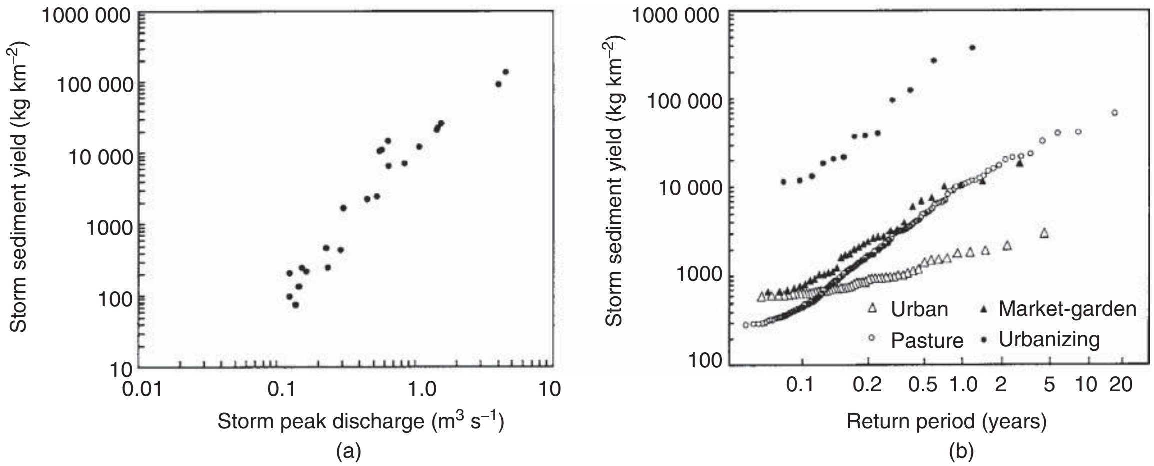


Figure 15.6 (a): Relation between suspended sediment yield and peak water discharge during storm runoff, Alexandra Stream, Auckland, New Zealand. (b) Magnitude–frequency relations for storm sediment yields at four small basins under different land uses, Auckland, New Zealand. After Hicks (1994). The return period is scaled in terms of the Extreme Value distribution.

in reservoirs, lakes and backwaters. Second, it influences the capacity of the sediment to adsorb and transport contaminants such as heavy metals. Finer sediment fractions (e.g. clays) tend to constitute platy silicate minerals that have larger specific surface areas and exhibit greater cation-exchange capacity (Ongley *et al.* 1982, 1990; Horowitz 1985; Walling and Moorehead 1989).

Typically, the type of particle size information required and hence the method of analysis vary with the application. Where the size data are required for hydraulic analyses, such as to predict settling in a reservoir or dispersion within a stream, a fall speed-based measure is required. However, if the actual physical dimension of the sediment grains is important, such as when considering sediment effects on fish, machinery, turbidity and acoustic properties and contaminant adsorption, then a direct measurement is required. Sometimes, both fall speed and physical size data are pertinent, as, for example, where contaminant-carrying sediment must be removed from a waterway by settling. Several texts (e.g. Guy 1969; Vanoni 1975; Allen 1990) detail size analysis techniques; a brief overview follows.

Most fall speed-based size analysis methods involve monitoring changes in sediment concentration or accumulation with time at a point in a settling chamber. Depending on the method, the sample is either introduced at the top of the chamber or is thoroughly mixed in the chamber before settling starts. The resultant data yield settling rates, which are converted into frequency distributions by mass of equivalent diameters of spherical quartz grains. Traditionally, the most common method has been the pipette method, wherein samples are extracted with a pipette (Krumbein and Pettijohn 1938). A variant is the bottom-withdrawal tube, wherein the accumulated sediment in a neck at the base of the chamber is withdrawn. Its advantage is that a smaller mass of sediment is required (0.5 g), but it is limited in its ability to resolve sand sizes (Guy 1969). Modern instruments automate these manual methods. For example, the SediGraph (Coakley and Syvitski 1991) uses X-rays to monitor sediment concentration in a settling chamber, whereas the rapid sediment analyser (RSA) (De Lange *et al.* 1997) records the weight of sediment accumulating on a pan at the chamber base. Both the RSA and the visual accumulation (VA) tube (Guy 1969) are designed for sand-size fractions. Since fall-speed analyses usually require a sediment mass of one to several grams (Vanoni 1975), the volume of sample that needs to be taken from the river to provide this mass will depend on the suspended sediment concentration. Porterfield (1972) provided nomographs for estimating sampling requirements for particle size analysis. Hydrometers are rarely used for suspended sediment analysis, owing to the relatively large masses of sediment required.

Several techniques directly measure grain dimensions. The traditional technique is sieving. Dry sieving is suitable for fine sand and coarser sediment (>125 μm). Its utility rapidly collapses for finer sediment owing to problems with sieve pore clogging and the effect of air currents within the vibrating sieve stack, which hinder the settling of fine sediment (M. Church, personal communication). Such problems are diminished by wet sieving, which is useful into the coarse silt range. Air-jet and sonic sieving are adjuncts to standard mechanical dry sieving analysis (Malhotra 1967). Laser diffraction spectroscopy is based on the principle that particles of a given size diffract light through a given angle, which increases with decreasing particle size (Agrawal *et al.* 1991). Laser back-scatter devices (e.g. Phillips and Walling 1995a; Galai Production 1997) record the time required for a laser beam to traverse an arc across individual particles. Particle size may also potentially be determined using phase Doppler anemometry (Cioffi and Gallerano 1991; Bennett and Best 1995). Analysis of digital imagery informs on both particle shape and size, can identify flocculation and may potentially be used to identify constituent minerals from their shape signature. We stress, however, that, except for quartz spheres, the different assumptions and approaches used with these techniques do not yield exactly the same result as a settling analysis.

Therefore, comparisons between the different techniques should be examined critically (McCave and Syvitski 1991).

A key advantage of most of the modern instrument-based methods is that they require only very small masses of sediment – an important consideration when dealing with dilute sediment concentrations. Also, some are portable (e.g. they can be set up on the stream bank) and some can even be deployed in situ (e.g. Bale and Morris 1991; Phillips and Walling 1995a; Gentien *et al.* 1995; Gray and Gartner 2009, 2010), providing information on both particle size and concentration.

A key decision to be made before undertaking suspended sediment particle size analysis is whether to break up (i.e. disperse) particle flocs. The difference between undispersed ('native' or 'effective') size distributions and dispersed (or 'ultimate') distributions is often substantial, with factor of 10 reductions in the median diameter being common (e.g. Walling and Moorehead 1989) (Fig. 15.7). In many applications, e.g. where the aim is to obtain fall-speed information, the sediment should be analysed as it occurs in the field. Traditionally, when only laboratory analysis was possible, samples were commonly duplicated or split, with one being kept for 'native' analysis and the other dispersed by chemical means (usually with a solution of sodium hexametaphosphate or Calgon). Nowadays, ultrasonic vibration also provides an effective dispersing mechanism. The problem with the native sample analysis, however, is that the original floc distribution may be altered between the river and laboratory bench, particularly if samples are left to settle in their bottles for an extended period before analysis (Phillips and Walling 1995b). Therefore, where particle flocs are important, analysing in situ, or as near in situ as possible, is desirable.

Off-the-shelf instruments capable of measuring and logging suspended sediment grain size in situ are becoming more widely available and used. An example is the LISST (http://www.sequoiasci.com/product/lisst-100x/). In various models, the LISST uses laser diffraction to provide a continuous record of the size grading and concentration, can periodically trap a flow sample and undertake a settling analysis, and can sample isokinetically from a streamlined housing. Such instruments point to a future where suspended sediment load by size fraction can be routinely gauged or continuously monitored at a point without the need to extract samples for laboratory analysis.

Although less is known about variation in suspended sediment size grading than about bulk loading, existing information indicates that it can vary substantially at a site during floods and seasonally (Walling and Moorehead 1989). Phillips and Walling (1995a) note that the particle size characteristics of a single 'instantaneous' sample (such as collected over a few seconds by a point sampler) cannot be considered representative of the load over a flood. Thus, as with the bulk sediment load, the variability in particle size demands an appropriate sampling strategy and the first task is to identify the main factors causing the variability. A great advantage of in situ particle size sensors is that they can integrate temporal variation and provide on-the-fly calibration data for optical and ABS sensors.

Synoptic sampling

Synoptic sampling of suspended sediment can provide an inexpensive indication of relative sediment sources within a drainage basin. Single-stage samplers may be used to collect synoptic datasets of near-surface sediment concentrations. These are passive sampling devices that are plumbed so that they fill, then seal, when the stage exceeds a set level (Edwards and Glysson 1999). Typically, they are deployed in vertical arrays so that a series of samples can be collected on either the rising or falling stage of an event (e.g. Gray and Fisk 1992). Their simplicity and cheapness mean that they are relatively easy to deploy in numbers through an entire drainage basin.

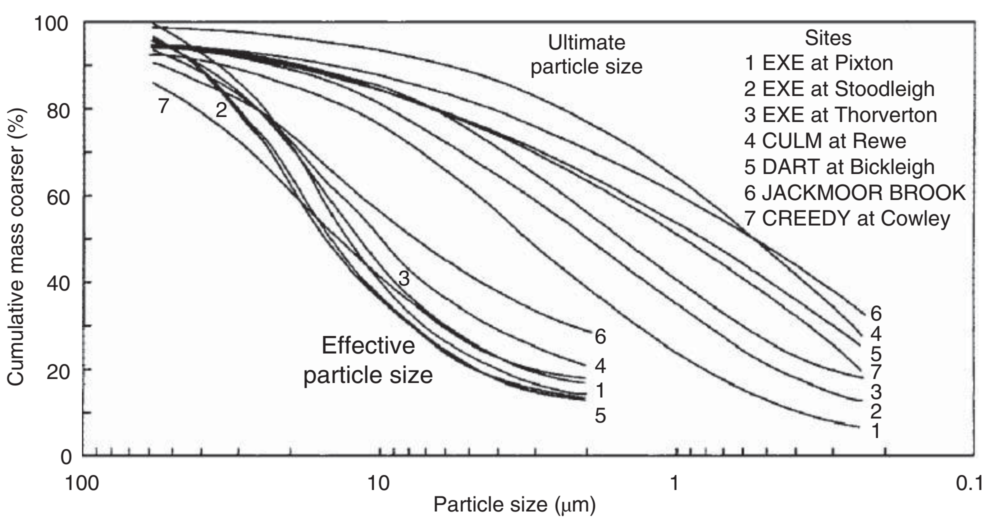


Figure 15.7 Comparison of native (effective) and dispersed (ultimate) size distributions of suspended sediment sampled at seven sites in the Exe River basin, England. Source: Walling and Moorehead, 1989. Reproduced with permission of Springer.

A synoptic perspective on suspended sediment concentrations in near-surface waters can also be gained from remotely sensed imagery (Gomez *et al.* 1995) (see also Chapter 6). Reflectance of the visible and near-infrared wavelengths is sensitive to scattering of radiation by fine-grained sediment particles. This permits near-surface suspended sediment concentrations to be determined from Thematic Mapper (TM) data after appropriate calibration using reflectance data for a range of native sediment–water mixtures (Witte *et al.* 1981; Ritchie and Cooper 1988; Mertes *et al.* 1993). We caution that such synoptic mapping of near-surface sediment concentration may not necessarily reflect the true distribution of sediment yield across a basin over an event, owing to the effects of different travel times and dispersion of sediment along tributaries, phase differences in rainfall and erosion processes across the basin and the absence of flow weighting.

15.4 Bedload sampling, measurement and prediction

Overview

From the inception of interest in bedload sampling in the 19th century (Humphreys and Abbott 1861; Davies 1900), the intent of most systematic surveys has been to characterize the bedload transport regime of the river in question by documenting the amount and size of sediment moved by different flows and to determine how these parameters change downstream (e.g. Swiss Federal Authority for Water Utilization 1939; McLean and Church 1999; McLean *et al.* 1999). Since the continuity of bedload transport is typically not maintained along a river, such information is crucial for identifying reaches in which the transport capacity will potentially be either larger or smaller than the supply and hence the conditions under which scour or fill of the river bed and other adjustments to channel geometry may occur (Ferguson *et al.* 2001; Ferguson and Church 2009; Gaeuman *et al.* 2009; Gomez *et al.* 2009). Restoration or enhancement of a river corridor in a manner that actively promotes the formation of viable aquatic habitats, the specification of channel maintenance or flushing flows and the effective in-stream management of sediment resources also require quantitative knowledge of bedload transport (American Society of Civil Engineers Task Committee 1992; Andrews and Nankervis 1995; Kondolf and Wilcock 1996; Wilcock *et al.* 1996; Schmidt *et al.* 2001; Czuba *et al.* 2011).

As with suspended load, the information needs and approach adopted to quantify bedload will vary with the situation. For example, they depend on whether information is required on bedload transport rates past a single section or whether a broader scale view is required, such as the average pattern of aggradation or degradation within a reach. At this juncture, it is also important to note that systematic bedload sampling is a time-consuming and expensive undertaking that currently is neither facilitated by commonly accepted equipment nor governed by a consistent set of protocols, even as increasingly comprehensive manuals become available on-line (see, for example, http://pubs.usgs.gov/twri/twri3-c2/html/pdf.html and http://www.whycos.org/whycos/sites/default/files/public/pdf/948_e.pdf). Moreover, a practicable solution to the complex problem of predicting bedload transport also remains tantalizingly out of reach, although significant advances in understanding have been made in the 21st century (Gomez 2006; Cui 2007; Turowski *et al.* 2011). For these reasons, we provide an overview of the available equipment and procedures rather than explicitly advocating the use of particular instruments and methodologies. We focus on describing two commonly utilized approaches for gaining knowledge of the bedload transfer through a reach of a river under a range of flow conditions:

1 field sampling or measurement;
2 application of an equation.

In this context, 'sampling' involves collecting discrete samples of bedload across a channel section for limited intervals of time, while 'measurements' involve recording the continuous or time-integrated bedload over the whole cross-section or reach by way of a trap, repeat surveys of riverbed morphology or monitoring the movement of individual particles (tracer).

Other surrogate methods for monitoring bedload that involve the use of active or passive sensors, such as acoustic Doppler current profilers (ADCPs) and geophones, continue to be developed and refined (Barton *et al.* 2010; Gray *et al.* 2010), but only methods and techniques for investigating bedload that have been broadly accepted for use and do not require elaborate installation are discussed here (Randle *et al.* 2009; Mao *et al.* 2010). For this reason also, physical model studies, the application of which is constrained to specific cases and the utility of which is exemplified by the Froude-scaled models of the braided North Branch of the Ashburton and meandering Sacramento rivers (Young and Davies 1990; Woidt *et al.* 2011) and Wallerstein *et al*'s (2001) distorted model of Abiaca Creek, are also not addressed here.

Bedload sampling

Sampler types

Many of the earliest sampling devices were of the basket type patterned on Mülhofer's (1933) design. These samplers retain sediment primarily by filtering it from the flow, but also because there is a reduction in flow velocity within the sampler. The minimum size of particle retained is determined by the mesh size of the basket, which is not intended to retain sand. Basket-type samplers typically have a large capacity, are designed specifically to accommodate very coarse gravel particles and continue to be used to sample gravel bedload (Bunte et al. 2004) (http://water.usgs.gov/fisp/docs/FISP_Tech_Memo_2009-1_Bedload_Traps.pdf).

Pressure-difference samplers were developed originally for use in sand-bed channels (Schaank 1937), where they continue to be deployed (Van Rijn and Gaweesh 1992; Gaweesh and Van

Rijn 1994), but they have also been developed for gravel-bed rivers (Helley and Smith 1971). The essence of their design is that the entrance velocity and the ambient flow velocity are equal. This is accomplished by constructing the sampler with walls that diverge towards the rear, creating a pressure drop at the sampler exit. Sediment is retained in a mesh bag mounted behind the sampler. Helley and Smith's (1971) simple design (Fig. 15.3a) has proven attractive, and Helley–Smith-type samplers (Childers *et al.* 1989; Ryan and Troendle 1997; Ryan and Porth 1999; Bunte and Abt 2009) are in common use. They range in size from lightweight versions with a 7.6 cm orifice, mounted on a wading rod, to heavier versions with a 15 cm orifice, deployed by cable from a crane on a bridge or from a cable-car.

Sampling and hydraulic efficiency

The presence of a sampling device on the river bed necessarily alters the pattern of the flow and sediment transport in its vicinity. Therefore, bedload samplers must be calibrated to determine their efficiency under different hydraulic and sediment transport conditions (Einstein 1937). Hydraulic efficiency is a measure of the degree to which the flow is accelerated or retarded by a sampler, defined as the ratio of the mean velocity of the flow through the sampler entrance to the mean velocity of the flow through the area occupied by the sampler entrance in the absence of the sampler (Hubbell 1964). Sampling efficiency indicates the extent to which a sampler either over- or under-samples the material in transport, defined as the ratio of the mass of the bedload collected during a specified time period to the true mass of bedload that would have passed through the entrance width in the same time period in the absence of the sampler (Hubbell 1964).

In practice, determining the hydraulic efficiency of a bedload sampler has proved to be a relatively simple task (Druffel *et al.* 1976), whereas determining the sampling efficiency is considerably more complex (Hubbell *et al.* 1985). Two factors confound the issue. First, the true bedload transport rate may only be defined indirectly (Hubbell *et al.* 1981). For example, transport rates typically vary with time and from point to point along and across a channel (Gomez 1983). Therefore, strictly, it is not possible to compare directly the transport rates determined at different points, irrespective of how close together the two points are located. Second, to obtain a single, constant measure of sampler efficiency directly, via a comparison of the mean at-a-point bedload transport rate, b, with measurements of the true mean bedload transport rate, t, the relation between b and t must be linear (De Vries 1973). Typically this has proved not to be the case (Einstein 1948; Hubbell 1987; Thomas and Lewis 1993b). Indeed, the sampler calibration process remains incomplete, not least because, even if sampler performance is optimized (Druffel *et al.* 1976; Johnson *et al.* 1977; Beschta 1981; Childers 1991; Gomez *et al.* 1991; Bunte and Abt 2005), inter-sampler transport relationships vary moderately with transport rate and among particle size classes and widely among streams (Emmett 1980; Pitlick 1988; Gaudet *et al.* 1994; Ryan and Troendle 1997; Childers 1999; Ryan and Porth 1999; Sterling and Church 2002; Ryan *et al.* 2005; Vericat *et al.* 2006; Bunte *et al.* 2008). Consequently, adjustment functions have been developed to align transport rates measured by Helley–Smith samplers with those measured with what are presumed to be more reliable devices (Bunte and Abt 2009).

Bedload sampling strategy and practice

Since bedload transport rates vary across channel and with time, appropriate temporal and spatial sampling strategies are required to minimize error in the estimates of the mean bedload transport rate. Temporal strategies involve three elements: the sampling time (the length of time the sampler remains on the bed), the sampling interval (the length of time that elapses between consecutive samples) and the sampling period (the sum of the sampling times and sampling intervals).

Any number of random samples may, in theory, provide an estimate of the prevailing mean bedload transport rate, although the magnitude of the errors involved may be expected to decrease as the sample size increases (Csoma 1973; De Vries 1973; Carey and Hubbell 1986). However, it is almost impossible to obtain a truly independent random sequence of bedload samples under field conditions and sequential samples likely will be serially correlated. Each observation in an auto-correlated time series repeats part of the information contained in previous observations, hence more sequential samples than independent random samples are required to provide the same information about the true mean. Nesper's (1937) experience prompted him to comment that, at 'low' transport rates, between 10 and 15 sequential samples were probably required to provide an acceptable indication of the mean transport rate, whereas about 30 samples were required at 'high' transport rates. Gomez *et al.* (1990) evaluated errors associated with at-a-point sampling where bedforms (dunes) were present. Their analysis suggests that 21 sequential samples are required to obtain an estimate of the mean at-a-point bedload transport rate that falls within 50% of the true mean rate at the 99% confidence level. This assumes that the sampling period is long enough to allow at least one primary bedform to migrate past the sampling point. Effects due to non-stationarity may be minimized by ensuring that the sampling interval does not coincide with the period of the bedforms that are present.

Note that the preceding discussion properly refers to the case of fully established motion, where the entire bed locally is taking part in the transport process. This is typically the case in sand and mixed sand and gravel bed rivers but may not be true in gravel bed rivers where conditions are near the threshold of motion and transport normally occurs at low rates (Andrews 1994; Buffington and Montgomery 1997; Church and Hassan 2005) and partial transport (i.e. where only the finer fractions on a graded gravel-bed surface are mobile) may be a commonly occurring condition (Wilcock and McArdell 1997; Hassan and Church 2000; Haschenburger and Wilcock 2003). In such cases, the transport is highly variable even in the short term, so serial correlation is apt to be low between successive samples and many samples are still required to characterize the variability. The sampling time will dictate the actual number of samples; however, if this time is protracted, it may absorb much of the short-term variability.

Reliable estimates of the streamwide bedload discharge obtained using sampling devices are dependent upon good at-a-point knowledge across the full width of the channel. The statistics of the sediment transport regime provide information on the number of times the sediment transport across the channel must be sampled in order to obtain a reliable value for the time-averaged bedload transport rate that conforms to reasonable limits (Kuhnle 1998). Gomez and Troutman (1997) showed that sampling errors decrease as the number of samples collected increases and the number of traverses of the channel over which the samples are collected increases. Assuming that sampling is conducted at a pace that allows a number of bedforms to pass through the sampling cross-section, bedload sampling schemes typically should involve four or five traverses of a river and the collection of 20–40 samples at a rate of five or six per hour. The objective is to reduce both random and systematic errors, and hence minimize the total error involved in the sampling process, by ensuring that spatial and temporal variability in the transport process is addressed.

Regardless of the manner in which the computational exercise is performed, the message is clear: the collection of reliable bedload data is a time-consuming process since the sampling period will, of necessity, be lengthy because sampling durations must be long enough to smooth out fluctuations (Gomez *et al.* 1989; Kuhnle 1996; Wilcock 2001; Singh *et al.* 2009; Fienberg *et al.* 2010). A corollary of this is that the sampling time may also be longer than is practicable with conventional sampling devices, which do not have the capacity to accommodate large amounts of sediment. For this reason, a sampling time that is of the order of 30 s typically is used. Since flow unsteadiness affects the rate of bedload transport, it is also assumed that the flow remains steady for the duration of sampling. In practice, since many rivers respond rapidly to precipitation inputs, this may prove an untenable requirement in all but snowmelt-dominated runoff regimes.

Care should be taken when using the sediment from a bedload sampler to characterize the size distribution of the bedload. This is because even the composite of a series of samples may be smaller than the minimum weight required to avoid bias and to achieve good precision in the calculated grain size percentiles (Ferguson and Paola 1997). A final practical comment is that great care should be exercised when raising and lowering bedload samplers, particularly when they are deployed with a cable. Over-sampling will occur if the sampler is allowed to 'shovel' into the bed or the face of a dune, and under-sampling can result if the sampler is not aligned flush with the bed surface, is allowed to rotate downstream or is tipped downwards during retrieval (Bunte and Abt 2009).

Bedload traps

Unlike the data obtained from bedload samplers, data obtained from bedload traps (devices installed below the bed surface) are usually regarded as exact (Church *et al.* 1991; Sterling and Church 2002). A trap is a cavity sunk into the streambed, with its upstream lip flush with the surface (Fig. 15.3c and d). The bedload falls into the trap and is retained in the cavity. Assuming that overfilling is not a problem, trap efficiencies of the order of 100% are to be expected if the opening is wide enough to prevent overpassing of saltating particles (Poreh *et al.* 1970; Habersack *et al.* 1998; Hassan and Church 2001; Sterling and Church 2002). Traps also have a distinct advantage over samplers, namely that if the trap spans the entire width of the river, it is possible not only to catch all the bedload that passes through the measuring section in a given period of time but also to measure continuously the rate at which sediment accumulates.

The simplest traps consist of lined pits or slots in the streambed in which the bedload collects over one or more events (Church *et al.* 1991). The bedload yield for the period in question is determined either volumetrically by surveying the deposit or manually by excavating and weighing the sediment (Hansen 1973; Newson 1980). More sophisticated traps incorporate pressure sensors that continuously weigh the mass of sediment in situ (Reid *et al.* 1980) or use a pump or conveyor belt to transfer it to a weighing station on the streambank (Dobson and Johnson 1940; Einstein 1944; Leopold and Emmett 1997). Other traps are designed so as to generate a vortex that ejects the sediment as it accumulates (Parshall 1952; Robinson 1962; Milhous 1973; Hayward and Sutherland 1974; Tacconi and Billi 1987) or separate the bedload from the fine sediment and water (Lenzi *et al.* 1999; Mao *et al.* 2010). Provided that they are not overfilled, traps invariably provide reliable data, but the limiting factor in their deployment is that they are often difficult and expensive to install.

Bedload tracer (see Chapter 14)

Tracer particles may provide an alternative or useful adjunct to the use of sampling devices or traps for measuring bedload transport rate. As discussed in detail in Chapter 14, tracer particles may be distinguished by painting, inserting magnets, transponders or transmitters or by utilizing inherent (or enhanced) natural properties (e.g. colour, magnetism). The distance travelled by tracer particles over a monitored epoch (e.g. a flood) can be used to calculate transport rate, but defining a general relation between transport rate and displacement length (or the virtual velocity of sediment) has proven to be a challenging task (Stelczer 1981; Wilcock 1997b, Haschenburger and Church 1998; Nikora *et al.* 2002; Ganti *et al.* 2010; Holmes 2010).

Morphological methods

The emphasis of many of the above techniques is on providing short-term data. However, at the event or intra-event scale, the most pronounced feature of the bedload transport process is its spatial and temporal variability (Gomez 1991; Ferguson and

Ashworth 1992). This variability reflects both variations in the flow conditions (i.e. transport capacity) and variations in the supply or availability of bed material. Factors influencing the bed material supply include event magnitude, the translation and dispersion of sediment waves (including bedforms such as gravel sheets), the presence of an armour layer and the occurrence of patches (Gomez *et al*. 1989; Parker 1990; Seal *et al*. 1993; Lisle 1995; Benda and Dunne 1997; Garcia *et al*. 1999; Lenzi *et al*. 1999; Lisle *et al*. 2000; Cui and Parker 2005; Cui *et al*. 2005; James 2010). From the perspective of characterizing the bedload transport regime in a particular reach, this spatial and temporal variability of bedload transport may be sufficiently complicated that it requires many measurements to reduce the variance in the observed data to an acceptable level. For this reason, the direct collection of quantitative data may always be an impractical method of estimating bed material transport rates in large rivers.

'Morphological' methods offer an alternative approach for determining bedload discharge. In essence, these are based on the continuity relation for bedload transport, which requires that the rate of change in the mean level of a segment of river bed is proportional to the difference between the transport in and out of the segment. Knowledge of the bed level change and the transport across one segment boundary permits computation of the transport rate across the other boundary. Natural situations arise where this relation can be exploited, at scales ranging from bed forms, to morphological units, to reaches.

Attempts have been made to determine transport rates from bedform statistics by comparing sequential bed profiles (Simons *et al*. 1965; Willis and Kennedy 1977). The geometry and movement of bedforms are highly variable, dunes are imperfect sediment traps and few transport measurements are available to validate results (Moll *et al*. 1987; Gabel 1993; Mohrig and Smith 1996; Nikora and Hicks 1997; Gaeuman and Jacobson 2007). Nonetheless, advances in sounding technology (such as integrated GPS and multi-beam sonar) that allow rapid and accurate collection of bathymetric data make this a viable alternative to direct sampling or measurement, at least in large, navigable rivers with active bedforms of adequate relief (Gaeuman and Jacobson 2007; Nittrouer *et al*. 2008; Abraham *et al*. 2011).

Neill (1971, 1987) developed an approach for defining the relation between morphological change and bed material transport in a systematically migrating meander bend. To derive a transport estimate, knowledge of the volume of sediment mobilized per unit length of channel and the average distance of travel (approximated as half the meander wavelength) is required. Church *et al*. (1986) described a more generalized approach in which knowledge of the changes in the volume of sediment stored in a reach and an estimate of the transport at one section permit the transport to be estimated throughout the reach. Carson and Griffiths (1989) used a similar approach to estimate gravel transport during flood in the large, braided Waimakariri River, New Zealand. McLean and Church (1999) compared two approaches that were used to estimate the annual gravel load of the lower Fraser River. The assumptions, procedures and limitations involved in the latter approach have also been discussed by Martin and Church (1995), who used it to estimate bed material transport in an 8 km long reach of the Vedder River.

Inasmuch as it yields information of quality comparable to that of direct measurements and requires less field effort, the morphological approach is relatively robust (Martin and Church 1995) and repeat surface or aerial surveys are now increasingly being used to estimate bedload transport (Lane *et al* 1995, 2003; Fuller *et al*. 2003; Rumsby *et al*. 2008). The procedure involves using the digital elevation models (DEMs) constructed after each survey to generate a DEM of difference (DoD) (Fig. 15.8a). Nominal accuracies of the survey methods currently in use, such as terrestrial laser scanning, rtkGPS (real time kinematic global positioning systems), LiDAR (light detection and ranging), multi-beam echo-sounding and ADP (acoustic Doppler profiling), typically are in the range ±several mm to ±0.2 m. These techniques are capable of rapidly generating very high spatial resolution point clouds, but in order to distinguish real geomorphic change in rivers from measurement error it is necessary to provide a robust, spatially variable estimate of DoD uncertainty. The usual approach for capturing this uncertainty in sediment budget calculations is to define a threshold above which elevation changes can be believed at a specified level of confidence (Fig. 15.8b). The threshold is based on measurement uncertainties from the component DEMs, but it may also be locally adjusted across the DoD using 'fuzzy inference' methods and/or the spatial coherence of erosion and deposition (Wheaton *et al*. 2010).

Bedload equations

Equations predict bedload transport capacity under given flow conditions. Their ability to do this is predicated on the assumption that it is possible to equate the rate at which bedload is transported to a specific set of hydraulic and sedimentological variables. Indeed, the underlying physics appears fairly straightforward (Du Boys 1879; Bagnold 1966), although it has been argued that a complete understanding of bedload transport requires that grain–grain (in addition to grain–fluid) interactions be fully accommodated (Frey and Church 2011). Ignoring the problems caused by variations in the supply or availability of sediment, it has long been recognized that, even at a constant discharge, bedload transport rates fluctuate (Fig. 15.9a). It is also apparent that if many observations are made over a period that is long enough to delimit the entire range of transport rates, a reliable estimate of the mean rate can be obtained (Einstein 1937). Consequently, field (and laboratory) data that are integrated over lengthy periods and across the whole width of the channel often yield coherent relations (Fig. 15.9b and c). The availability of such data seemed to confirm the existence of a bedload function and to demonstrate that bedload transport indeed occurred in accordance with established principles (Müller 1937). This, coupled with the realization that sampler calibration was by no means a straightforward task (Einstein

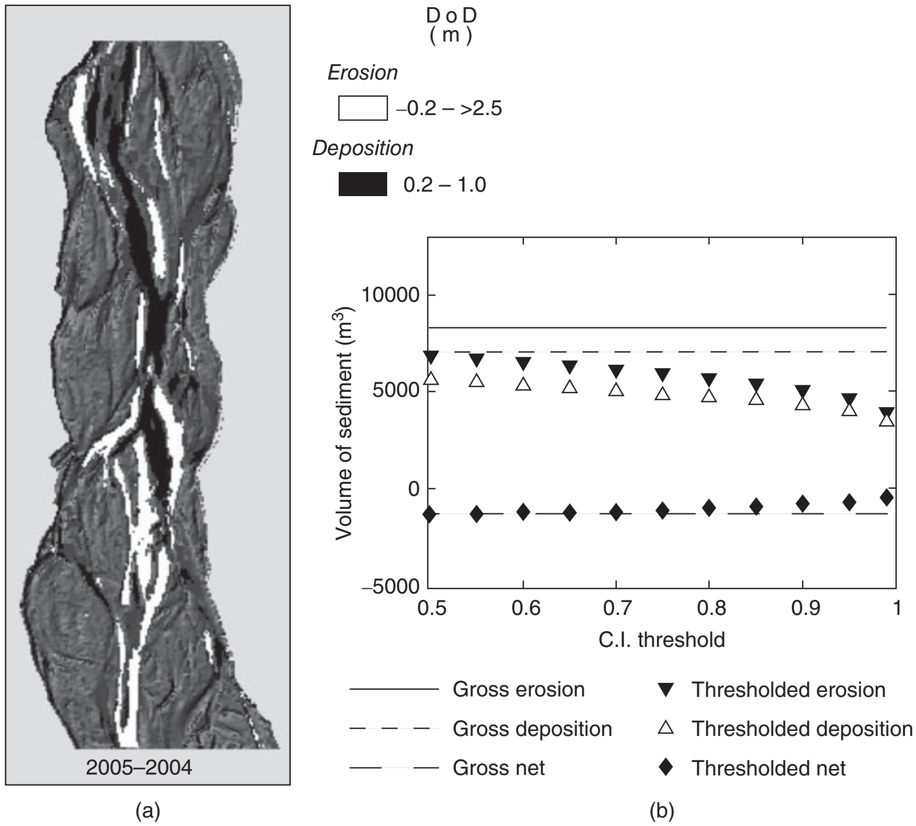


Figure 15.8 (a) Map of erosion and deposition for a 1 km reach in the River Feshie, UK, over the period 2004–2005, derived from thresholded DEM of difference (DoD) at 95% confidence interval, with adjustment for spatial coherence of erosion and deposition. (b) Sensitivity of sediment budget estimates to the confidence interval (CI) threshold used. Source: Wheaton *et al*, 2010. Reproduced with permission of Wiley.

1937; Nesper 1937), helped foster the view that the prediction of bedload discharge was a viable proposition. This may indeed be the case in rivers where there is a high availability of sediment in relation to runoff, under conditions where the bed is fully mobile (Gomez 2006). In such rivers, there is an upper limit on bedload transport rates governed by the efficiency of energy expenditure and a straightforward relation between bedload transport efficiency and the median particle size of the bed load emerges (Fig. 15.9d). In practice, however, the complexities of the interrelations between the conditions governing bedload transport confound the issue (Gilbert 1914).

Most equations describe a relation that has been either defined empirically on the basis of laboratory or field data or derived from basic mechanical or physical principles. From the outset, two issues have contributed to the profusion of bedload transport equations. First, there is no consensus about the fundamental hydraulic and sedimentological quantities involved. Second, dissatisfaction with the performance of a particular equation (which was often inspired by its poor performance against data that were not included in the initial analysis) encouraged attempts to develop new relations. To accommodate different levels of transport intensity, the most widely utilized of these new relations comprise more than one function. They include Parker *et al.*'s (1982) and Parker's (1990) substrate- and surface-based equations, Wilcock's (2001) surface-based two-fraction equation, Wilcock and Crowe's (2003) surface-based relation and Cui's (2007) unified gravel-sand model. The Wilcock and Crowe relation, derived from laboratory studies, has since been modified slightly using field data by Gaeuman *et al* (2009). Several such and other well-known relations have been incorporated in a spreadsheet-based program designed to assist with the prediction of bedload transport in gravel-bed rivers (Pitlick *et al.* 2009) (http://www.fs.fed.us/rm/pubs/rmrs_gtr223.html). Monte Carlo analysis may be used to assess the uncertainty associated with transport estimates, which can be calibrated using bedload samples (Wilcock 2001; Wilcock *et al.* 2009). However, it remains a source of some discomfort that there appear to be more bedload equations than there are reliable data sets by which to test them (Gomez and Church 1989) and field testing of equations over long (decadal) time-scales has rarely been undertaken (Martin 2003). At this scale, recourse is typically made to essentially qualitative comparisons based on trends in the available data (Ferguson and Church 2009; Gomez *et al.* 2009).

There have been a number of major reviews that use field data to compare bedload transport equations (Table 15.1). None have

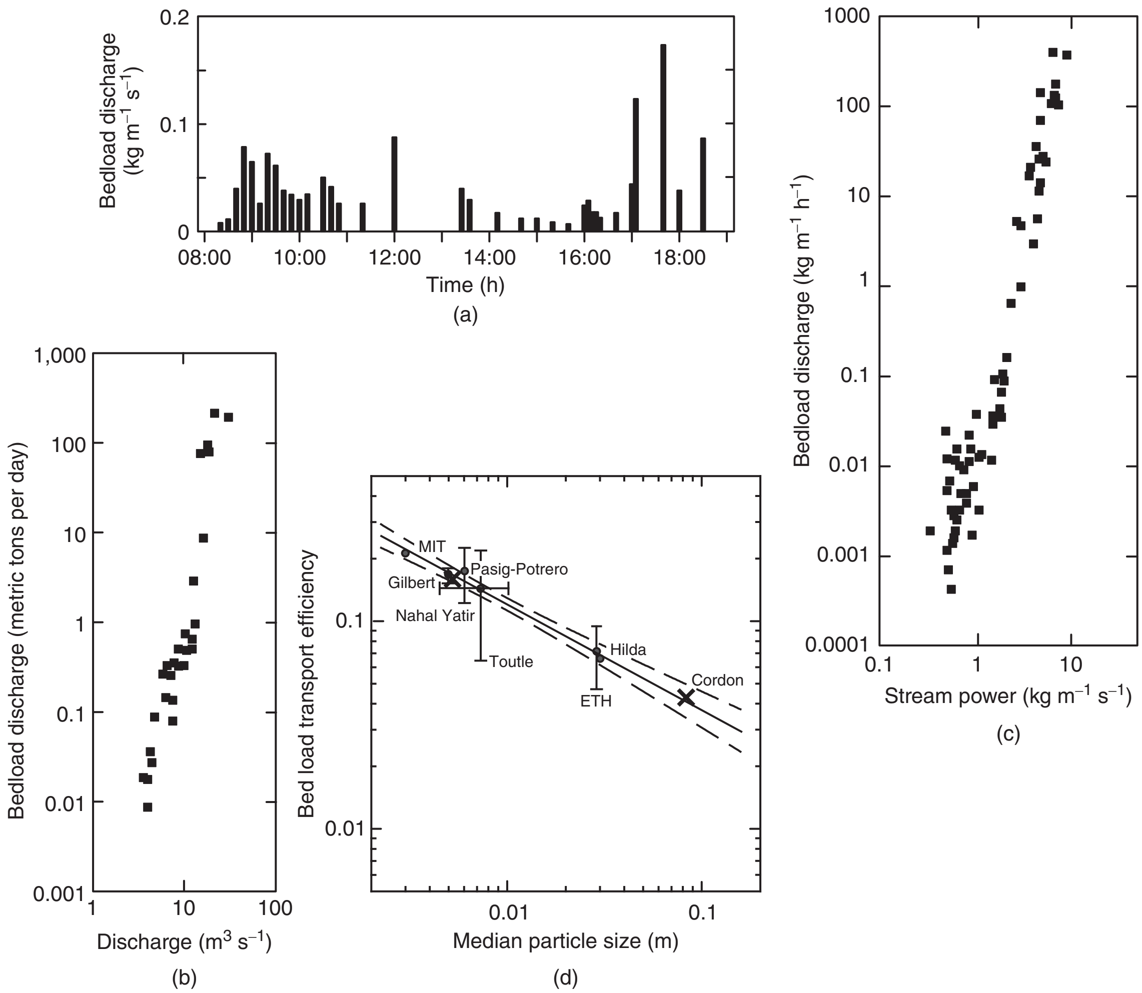


Figure 15.9 (a) Temporal variations in bedload transport rates observed at virtually constant discharge (0.62–0.63 m^3 s^{-1}) in Torlesse Stream, New Zealand, 30 August 1973. Data from Hayward (1980). (b) Observed relation between bedload transport rate and water discharge in the Enoree River, South Carolina, 1939–1940. Data from Dobson and Johnson (1940). (c) Observed relation between bedload transport rate and stream power in Oak Creek, Oregon, 1971. Data from Milhous (1973). Note that all of these data were obtained using pit traps that spanned the entire channel width. (d) Relation of average bedload transport efficiency ($i_b \tan a/w$, where i_b is immersed-weight unit bedload flux, w is unit stream power and a is friction angle) to bedload median particle size (D_{50}) for conditions in gravel-bed rivers and flumes where there are no constraints on sediment supply or availability. Error bars indicate standard deviation of bedload transport efficiency and median particle size of bedload (when reported). The regression relation (solid line, $r^2 = 0.99$; dashed lines denote the 95% confidence intervals) suggests the relation $i_b = w(0.0115D_{50} - 0.51)/0.63$ applies to such conditions. Crosses denote independent data used to test the relation's applicability. Data from Gomez (2006).

provided definitive results. In consequence, no single equation, or even a small group of equations, has been universally accepted or recognized as being especially appropriate for practical application and it is important to remember this point when applying numerical models that simulate river channel morphological change. In the face of such overwhelming indecision, an interested party has little option but to make an intuitive selection on the basis of the similarity of the conditions for which a particular equation was derived and those in the river in question. The indexes developed by Williams and Julien (1989) and Bechteler and Maurer (1991) may assist with this process. However, since there is no reason to suppose that any equation will necessarily provide complete correlation, caution dictates that the results of several equations be compared. Indeed, in the case of morphological modelling, the calibration process often involves adjusting equation coefficients to permit a match between modelled and observed bed-level changes.

Most errors in equation application arise from the input data (Wilcock *et al.* 2009). In applying any one-dimensional equilibrium transport equations, local hydraulic parameters (as against section-averaged parameters) should be utilized as the use of average values represents a channel-wide integration before the transport calculation. The effect may be important since most equations are non-linear. However, there are obviously problems involved in maintaining strict observance of the form of any equation, not least, for example, because local bed shear

Table 15.1 Summary of major reviews of bedload transport equations.

Study	Equations examined	Methodology	Recommended, representative or preferred equations
Vanoni *et al.* (1961)	Du Boys–Straub, Einstein, Einstein–Brown, *Laursen, Meyer–Peter, Meyer–Peter and Müller, Schoklitsch 1934, Shields	Comparison of sediment rating curves	None specified
Shulits and Hill (1968)	Du Boys–Straub, Casey, Einstein, Elzerman–Frijlink, Haywood, Kalinske, *Larsen, Meyer–Peter, Meyer–Peter and Müller, Rottner, Schoklitsch 1934, Schoklitsch 1943, Shields, USWES	Determination of limits of agreement between calculated bedload transport rates	Du Boys–Straub, Meyer–Peter and Müller, Schoklitsch 1934
ASCE Task Committee (1971)	Blench, Du Boys–Straub, *Colby, Einstein, Einstein–Brown, *Engelund–Hansen, *Inglis–Lacey, * Larsen, Meyer–Peter, Meyer–Peter and Müller, Schoklitsch 1934, Shields, *Toffaleti	Comparison of sediment rating curves	*Colby, *Engelund –Hansen, *Toffaleti
White *et al.* (1973)	*Ackers–White, Bagnold 1956, *, Bagnold 1966, Bishop, *Simons–Richardson, *Blench, *Einstein, Einstein–Brown, *Engelund–Hansen, *Graf, *Inglis, Kalinske, *Laursen, Meyer–Peter and Müller, Rottner, Shields, *Toffaleti, Yalin	Comparison of discrepancy ratios	*Ackers–White, * Engelund –Hansen, Rottner
Mahmood (1980)	*Ackers–White, *Colby, *modified Colby, Einstein, * modified Einstein, *Engelund–Hansen, *Laursen, Meyer–Peter and Müller, Mahmood, Shen–Hwang, *Toffaleti, *Yang	Comparison with the modified Einstein procedure	Shen–Hwang, *Toffaleti
Gomez and Church (1989)	Ackers–White, Ackers–White–Day, Ackers–White–Sutherland, Du Boys–Straub, Bagnold 1980, Einstein, Meyer–Peter, Meyer–Peter and Müller, Parker, Schoklitsch 1934, Schoklitsch 1943, Yalin 1963	Comparison of mean and local bias	Ackers–White–Day, Bagnold 1980, Einstein, Parker
Barry *et al.* (2004)	Eight variants of Meyer–Peter and Müller 1948, Ackers and White 1973 as modified by Day 1980, Bagnold 1980, Parker *et al.* 1982 equation as revised by Parker 1990 (subsurface-based version)	Paired sample $\chi 2$ test	Bedload transport best described as a simple power function of water discharge

*Total load equations.

stress cannot be measured directly and will vary widely across a gravel-bed river (Dietrich and Whiting 1989; Wilcock *et al.* 1994). In coarse-grained channels, the point at which motion is initiated may depend more on the relative size than the absolute size of the bed material. That is not to say that the fundamental effect of particle weight is eliminated, but rather that, because of effects due to sheltering, protrusion, grading and shape, it becomes less dominant. Starting with Einstein (1950), several models have sought to account for these factors (Parker *et al.* 1982; Parker 1990; Andrews and Smith 1992). More recently, Wilcock (1997a) sought to circumvent the problems caused by the need to provide details of local flow and sediment properties by developing a model that relies on suitably defined mean properties of the flow and sediment in a stream reach. Accounting for surface structure in gravels is also a significant obstacle to the assured application of any bedload equation in the field (e.g. Jackson and Beschta 1982; Lisle and Madej 1992; Seal *et al.* 1993; Hassan and Church 2000). Bed surface structure may reflect the flow history, and its creation or disruption is known to influence bedload transport rates (Gomez 1991). Moreover, because the summary effects of structure are not readily measured, assumptions about fundamental parameters, such as the Shields number, are not easily made.

The selection of an appropriate equation for use in sand-bed rivers, where relative size effects are not an issue and the constraints on the availability of sediment are relaxed, may be more straightforward. A variety of field data suggest that the unit discharge of sand varies approximately with the fifth power of mean velocity and inversely with particle diameter (Posada and Nordin 1993). Several theoretical relations, such as that developed by Engelund and Hansen (1967), conform with such a trend, although all incorporate one or more empirically derived coefficients.

Bedload rating curves

Although it requires verification with field data, the appeal of an equation is that it produces a rating that may, in principle, be used in conjunction with discharge data to compute bedload yield over a specified period (Emmett and Wolman 2001; Wilcock *et al.* 2009). Although there is often considerable scatter and the data rarely extend across the entire range of flow conditions, field data may also be used to define a rating curve. Simple functions are typically used (Wilcock *et al.* 1996; Moog and Whiting 1998), with the transport rate commonly portrayed as a power function of discharge (Barry *et al*. 2004, 2008). Irrespective of whether an equation, field data or a combination are used to construct a bedload rating curve, it is common practice to compute transport rates for the sand and gravel fractions separately (Wilcock 1998). This is because all particle sizes present on the bed are rarely in motion at once and the bedload size distribution only infrequently approaches that of the bed material (Gomez 1995; Lisle 1995).

15.5 Total load

Determining the total sediment load requires matching complementary methods of determining bedload and suspended load. It is important that this matching is done over time and spatial scales that are consistent with the component methods; also, double accounting of size fractions that overlap the suspended and bed loads should be avoided. Primarily, the approach for determining total sediment load depends on whether the stream bed material is sand or gravel.

For sand-bed streams, there are three total load approaches: equation, sampling and a combination. So-called 'total load' equations (e.g. Engelund and Hansen 1967; Toffaleti 1968) actually only determine the bed material load and they are appropriate where there is no washload and an unrestricted supply of sand from the bed. These require input data on flow hydraulics and bed material size characteristics. Where there is washload and/or restrictions on the sand supply, one approach is to sample the suspended load and to compute the suspended and bed loads in the unmeasured zone (Fig. 15.2), as in the 'modified Einstein' and related approaches (Colby and Hembree 1955; Stevens and Yang 1989). Alternatively, bedload and suspended load can both be sampled. Ideally, the sampling of both modes should be synchronous, using a device such as the Delft Nile sampler (Van Rijn and Gaweesh 1992; Gaweesh and Van Rijn 1994). This combines a bedload sampler specially designed for sand beds with a vertical array of pumping point samplers. Combined (although not synchronous) bedload and suspended load sampling over sand-beds has also been conducted using Helley–Smith-type bedload samplers and depth-integrating suspended sediment samplers (e.g. Andrews 1981). With this approach, it is necessary to correct for any double-accounting of sand fractions in the depth range intercepted by the bedload sampler. In a naturally contracted section or turbulence flume, sand bedload is forced into suspension and can be treated as suspended load. Colby and Hembree (1955) and Hubbell and Matejka (1959) employed this procedure to determine the total sediment load of rivers in the Nebraska Sand Hills.

The combined sampling approach is also an option for gravel-bed streams. Again, any double-accounting of size fractions intercepted by both the suspended and bed load samplers in the near-bed zone needs to be addressed. Alternatively, the bedload component can be determined using equations. The morphological method for 'bedload' actually determines the time-averaged bed material load; this needs to be combined with the washload component of the suspended load over the same time frame.

15.6 Estimating sediment yields from reservoir sedimentation

Reservoirs present special needs for sedimentation information (such as their rate of infilling and the rate of depletion of bed material load to the channel and coastline downstream), but they also afford unique and robust opportunities for measuring the total sediment load of the inflowing river(s) on event, inter-annual or long-term average bases. Because they are backwaters, reservoirs trap part of the washload of inflowing streams in addition to the bed material load (Brune 1953; Maneux *et al.* 2001), but the degree of entrapment of each size fraction depends on the hydraulic conditions through the reservoir, which vary with time. For this reason, it is often easier to determine a reservoir sediment budget, and the total inflowing sediment load, by combining reservoir sedimentation volumes with the suspended load in the outflow.

Techniques for surveying sedimentation volumes in reservoirs are well detailed in several texts (e.g. Vanoni 1975; Morris and Fan 1998). Typically, water depth is sounded by boat either with a weighted line or an echo sounder, while horizontal position may be measured from a tagline, a Total Station system or differential GPS (DGPS) (e.g. Schall and Fisher 1996). The most modern approach is to use DGPS with multi-beam or swath sonar, which provides spatial detail adequate to develop a digital elevation model (DEM) of the reservoir bed (e.g. US Bureau of Reclamation 2006). A DEM combined with readily available topographic/GIS software permits easy computation of volume changes and also mapping of sedimentation depths (e.g. Sullivan 1996). The interval between surveys will depend largely on the rate of sedimentation, but typically may be 5–10 years.

To reconcile sedimentation volumes with other sediment budget information (which is usually measured in units of mass flux), it is necessary to determine the bulk density (or specific weight) of the reservoir deposits. Measuring this is relatively easy if a reservoir is periodically dry, which facilitates the extraction of cores or in situ measurements. A simple method to use with either cores or in situ excavations is the 'sand cone' approach (e.g. Vanoni 1975), wherein the mass of sediment removed from a hole is weighed and the volume of the hole is determined by refilling it with sand. The density of submerged deposits may be measured in situ using probes that are pushed into the bed sediment, such as a gamma probe (e.g. McHenry 1971), measured in the laboratory from core samples or estimated based on analysis of the grainsize in sediment cores and an empirical relation between bulk density and grainsize of freshly deposited sediment (e.g. Vanoni 1975). The bulk density of reservoir sediment increases with time due to consolidation. Various semi-empirical approaches have been developed to estimate this increase in density or alternatively to estimate the settling of a sedimentary layer due to consolidation (e.g. Lane and Koelzer 1943; Miller 1953; Gill 1988).

Reservoir outflows, because their suspended loads are typically fine grained and well mixed, are usually well suited to continuous monitoring with turbidity sensors. Sediment ratings for reservoir outflows often show wide data scatter owing to phase lags between the water discharge and sediment concentration peaks and to artificial manipulation of the outflow discharge.

Stratigraphic and sedimentological analysis of reservoir deposits, sampled from cores or from excavations if the reservoir dries out, can provide detailed information on event sediment yields and long-term average yields (Laronne and Wilhelm 2001). With adequate dating control on the stratigraphic record, series of event sediment yields may then be transformed into a probability density function or a magnitude-frequency relation similar to that shown in Fig. 15.6b. With a record spanning many decades, long-term changes in the mean annual sediment yield or probability density function of event yields may be related to factors that influence catchment erosion such as land use or climate variability (Laronne 1990; Seydell 1998; Yeloff *et al.* 2005).

A caution is that reservoir catchment boundaries are sometimes confused by water diversion/distribution schemes, so that sediment yields calculated from reservoir sedimentation surveys may not relate directly to the local catchment characteristics (Butcher *et al.* 1992).

15.7 Key points for designing a sediment measurement programme – a summary

The basic steps in designing a sediment measurement programme are set out in Table 15.2. First it is necessary to define the purpose(s) of the programme, since this largely determines the basic measurement approach. With this decided, the measurement 'tools' can be selected with the aid of Tables 15.3 and 15.4. Some prior knowledge of the relative importance of suspended load and bedload and of the size grade of the bedload (whether sand or gravel) will help to focus the measurement effort and choice of tools.

The suspended load needs to be sampled, since typically it is limited by the supply of fine sediment to the channel rather than by physical transport capacity. A key consideration is whether the problem at hand requires near-continuous data, event-based information or simply long-term statistics such as the mean annual yield (Table 15.3). Continuous data require index or point samples collected manually or by auto-sampler or else surrogate records from optical or acoustic sensors. Such point measurements need to be related to the cross-section mean sediment concentration. If the sampling purpose is only to determine the average annual sediment yield, then the sediment rating or direct estimation methods are more economical alternatives to continuous monitoring. Suspended sediment ratings either attempt to model explicitly the sediment concentration as a function of all significant controlling factors or, more commonly, are used to model the conditional mean concentration over the period of interest as a function of water discharge. Care is required in fitting sediment ratings, in correcting for bias induced by data transformations and with sampling strategies for compiling rating datasets.

Suspended sediment yields for discrete runoff events can be related to indices such as event peak flow. Event-yield magnitude–frequency relations are useful for discriminating land-use effects on sediment supply. Synoptic sampling of near-surface waters, either with manual samples or with remote sensing, provides useful relative indicators of basin-wide sediment sources.

Suspended sediment particle size influences entrainment, mixing, deposition, downstream sorting and the capacity of sediment to adsorb and transport contaminants. Methods for determining particle size are based either on analysis of fall speed or on direct measurement of physical dimensions. The method used depends on the problem. A decision is required whether to determine the effective particle size distribution or the ultimate distribution, after particle flocs have been dispersed. In situ sensors avoid the problem of having the effective distribution change between stream and laboratory.

It is relatively easy to start a programme of suspended sediment monitoring; knowing when to stop it is less straightforward. Studies of long records of continuous sampling have shown that annual sediment yields, at least in small catchments, typically have an approximately log-normal distribution and high variability (Renard and Lane 1975; Van Sickle 1981). Day (1988) analysed long records (up to 30 years) from Canadian rivers and found that the mean characteristics of the suspended sediment yield stabilized (with a stable standard error of the mean) after approximately 10 years. Thus, assuming that no longer term trend or non-stationary signal exists, such as induced by land-use change or catastrophic climatic and tectonic events, a decadal time span for monitoring is suggested.

Bedload may be determined by field sampling and measurement or by equation (Table 15.4). The hydraulic and sampling efficiencies of bedload samplers vary with their design and the sampling efficiency is difficult to establish definitively. However, in practice such deficiencies are less important than having the correct sampling strategy to overcome the considerable spatial and temporal variations in bedload transport observed in rivers, particularly those with gravel beds. It may not always be possible to collect quantitative bedload measurements because of scale considerations. For example, there are practicable limits to the size of a river in which samplers can be deployed. Bedload traps are more exact devices, but are limited in size and are expensive, hence they are generally limited to research applications in narrow channels.

Table 15.2 Key steps and considerations when designing a sediment measurement programme.

No.	Step	Examples/considerations
1	Decide main purpose of measurements	Statistics of instantaneous sediment load Annual-average total sediment load Erosion/deposition in a river reach or reservoir Scientific study of fluvial processes in a river reach Other/a combination of the above
2	Identify nature of sediment load of primary interest	Suspended load, bedload or total load Expected suspended/bedload ratio Composition of bedload, e.g. sand or gravel
3	Decide basic temporal sampling approach appropriate to purpose determined in step 1	Continuous sampling Event-based measurements Statistical sampling to determine only annual average loads
4	Choose measuring approach to suit outcome of steps 1–3 and accuracy requirements	In situ sensors for suspended sediment concentration and bedload Manual samplers for bedload and suspended load Automatic suspended sediment samplers Bedload traps Surveys of erosion/deposition in river reaches or reservoirs Bedload tracers Bedload or total load equations Remote sensing of suspended load Merging bedload and suspended load measurements For bedload, use a combination of methods to reduce uncertainty
5	Select basin-scale spatial sampling strategy to suit purpose from step 1	Network of measurement stations in river basin Inflows/outflows of reach of interest Synoptic sampling
6	Design at-a-section spatial sampling strategy to suit measurement approach from step 4	Sampling verticals Point versus cross-section mean calibration relations
7	Design temporal sampling strategy to suit approaches decided in steps 3 and 4	Duration of discrete measurements (e.g. bedload samples) Time base for auto-sampling (e.g. fixed time, flow-proportional) Time interval between measurements/surveys Duration of measurement programme
8	Determine requirements for analysis of particle size	In situ or laboratory measurement Effective or ultimate size distribution of suspended load Adequate mass sampled for analysis technique
9	Identify supplementary data needs (e.g. for computing sediment discharge using rating relations or equations; for converting sediment volumes to masses)	Sediment mineral and/or bulk density Water discharge records Flow hydraulic data, e.g. channel geometry, slope, roughness

Morphological methods provide a reasonably robust estimate of the time- and space-averaged bedload, even on large rivers, provided that the field conditions are appropriate for the method and some independent means of confirming the transport estimate is available. Although there are many bedload equations, none has been universally accepted. They all apply more or less to a limited range of conditions and none reproduces the short-term fluctuations in bedload transport rates seen in nature.

Thus, except for traps, none of the existing methodologies for estimating bedload is inherently reliable. Indeed, Hubbell's observation that 'no single apparatus or procedure, whether theoretical or empirical, has been universally accepted as completely satisfactory for the determination of bedload discharge' (Hubbell 1964, p. 2) remains current. Hence caution dictates that a combination of techniques be used to estimate bedload discharge and their results compared. Carson and Griffiths (1987) and McLean and Church (1999) provide an indication as to how this might be done. End-users should also be aware that to address many environmental and management issues effectively, a more comprehensive (and inevitably longer term) perspective on sediment transfers within a basin is typically required than is provided by a site-specific characterization of a river's bedload transport regime.

Determining the total sediment load requires matching methods for determining suspended load and bedload. These should have consistent time bases and some correction may be required to avoid double-accounting of size fractions that appear in both the bedload and suspended load. Simple 'total load' equations are appropriate only for sand-bed channels lacking washload.

Reservoirs offer unique and robust opportunities for measuring the sediment load of their inflowing rivers, both on a long-term average basis, from periodic surveys of bed levels, and on an event basis, from sedimentological/stratigraphic analysis of their bed sediment.

Table 15.3 Tools, typical applications and constraints on information about suspended sediment (SS).

Information requirement	Application	Tools	Constraints
Instantaneous SS concentration or load	Determine cross-section mean SS load or discharge-weighted mean concentration under given, steady flow conditions	Point sampling to produce concentration and velocity profiles or depth-integrated sampling with matching water discharge gauging, all at multiple verticals	Time consuming
Continuous SS concentration or load	Continuous records of SS load, concentration or turbidity for determining statistics such as ranges, exceedance probabilities, mean, annual variability	Single-point index sampling with manual sampler, auto-sampler, optical or acoustic sensor	Requires relations calibrating point values to cross-section mean SS concentration
Long-term average SS concentration or load	Long-term average SS yield, e.g. for reservoir sediment inflows	*C* versus *Q* sediment rating combined with either flow duration table or flow time series; multivariate rating and appropriate time series data.	Accuracy limited by sampling strategy, number of samples, rating model fitting
		Direct load estimation using stratified or variable probability sampling strategies with data loggers and auto-samplers	Requires auto-sampler, data logger, calibration relations
Event-based SS concentration or load	Event sediment yields or peak concentrations, e.g. for predicting inflows to small reservoirs and water clarity in estuaries	Storm sediment yield ratings, reservoir stratigraphy, event-yield magnitude–frequency analysis	Need point to cross-section mean calibration relations when relying on continuous sampling; need bulk density measurements of reservoir sediments
Synoptic sampling	Mapping relative sediment sources across basins	Multi-spectral analysis of satellite/aerial imagery, manual sampling, single-stage samplers	Requires calibration of SS concentration to image signature, near-surface data only, synoptic map may not represent event average, due to phase differences in sediment supply and transport from tributaries
Particle size by settling analysis	Entrainment, mixing, deposition issues	Pipette, bottom-withdrawal tube, hydrometer, visual accumulation tube, rapid sediment analyser, sedigraph	Manual methods time consuming, requires minimum mass of sediment, sand and finer fractions analysed separately
Particle size by physical size analysis	Machinery damage, sediment filtering, contaminant adsorption and transport	Wet sieves, laser diffraction devices, laser back-scatter devices, microscopic image analysis	Non-standard measurements among devices, basic distributions often by grain count not by sediment mass, cannot be directly compared with settling analysis results
Effective/ultimate particle size	Adsorbed contaminant transport and management, sediment settling and water clarity issues	Chemical and dispersing agents and ultrasonic devices, in situ laser sensors	Undispersed sample properties may alter between sampling and laboratory

15.8 Case example: sediment budget for Upper Clutha River, New Zealand

The Clutha River drains 20,500 km^2 of mainly schist terrain in South Island, New Zealand, and has a mean flow near the coast of 565 m^3 s^{-1}. The upper river is used for hydro-electricity generation, with dams built at Roxburgh in 1957 and upstream at Clyde in 1992 (Fig. 15.10a). Most of the runoff into the upper Clutha is sourced from the wetter, northwest corner of the basin and passes through three large natural lakes. Sediment is derived from tributaries downstream of these lakes, particularly from the Shotover River, which has the largest (1088 km^2) and steepest catchment and receives the highest annual rainfall (> 2000 mm). Sediment loads in the upper Clutha system have been monitored since the 1960s, the main purposes being to quantify inputs to existing and planned hydro-reservoirs and, at least in the early years, to clarify sediment source areas to establish the practicality of reducing the sediment supply using soil conservation measures. The following summarizes this monitoring programme and the results obtained for the period up to 1992, when the Clyde Dam was commissioned.

Suspended sediment has been gauged (i.e. using depth-integrating samplers at multiple verticals) at flow recording sites on all of the major tributaries (Fig. 15.10a), with the aim of determining mean annual yields via the sediment rating approach. The most recent yield estimates, using LOWESS to fit ratings to

Table 15.4 Tools, typical applications and constraints on information about bedload.

Information requirement	Application	Tools	Constraints
'At-a-point' bedload transport rate	Characterization of temporal variability in bedload transport rates and estimation of mean rate under given, steady flow conditions. Determination of conditions for incipient motion of a given size fraction	Basket and pressure difference samplers; compartmentalized, continuously recording pit traps	Sampling: accuracy limited by number of samples obtained. Pit traps: expensive to construct
'Stream-wide' bedload transport rate	Estimation of mean bedload transport rate under given, steady flow conditions. Determination of bedload discharge. Characterization of patterns of scour and fill. Construction of i_b versus Q rating	Samplers and traps; bedform surveys; equations	Time consuming and labour intensive. Equations require calibration/verification with field data
Bedload yield	Estimation of bedload yield on an event, seasonal or multi-year basis	Traps; surveys of sedimentation basins or reservoirs; morphological methods; i_b versus Q rating; tracers and scour chains	Morphological methods require information on bedload across reach boundary. Rating curves require calibration with field data. Limited recovery of tracers

the log-transformed datasets, show a total suspended load from tributaries upstream of Lake Roxburgh of 1.97×10^6 t per year, with the Shotover River supplying 67% of this. For some gaugings at each site, duplicate samples were bulked and analysed for particle size (usually with a bottom-withdrawal tube), with the results averaged to estimate a representative size grading. The relations between suspended sediment concentration and water discharge are comparatively poor on the Kawarau and Clutha Rivers, since both receive much of their flows as clear water from the natural lakes; thus over the period 1977–1980, daily index samples were collected from sites that covered the inflows and outflows to Lake Roxburgh and the future Lake Dunstan Reservoir. Depth-integrated multi-vertical gaugings were used to develop relations between index sample concentration and cross-section mean concentration and also to measure the particle size of the inflowing and outflowing suspended loads. The results confirmed that the main source of sediment was from the Shotover River, via the Kawarau River, and showed that the average trap efficiency of suspended sediment entering Lake Roxburgh was 80% (Jowett and Hicks 1981).

Bedload was sampled in the Shotover River over a range of flows using a 150 mm wide orifice Helley–Smith sampler operated from a motorized cableway. Each bedload measurement involved repeat traverses of 20 verticals and all samples of the sandy gravel bedload were analysed for particle size (Fig. 15.10b). The gauged bedload discharges were used to verify (albeit with considerable data scatter) a bedload rating (Fig. 15.10b) derived using the approach of Wilcock (1997a, 1998), which involves different transport functions for the sand and gravel fractions and relates the threshold of motion of sand and gravel to their relative proportions. The bedload yield of the Shotover so estimated was 0.26×10^6 t per year, which is equivalent to 20% of the suspended load. Because of the cost of bedload sampling and because the Shotover was the dominant sediment source, the bedloads of the other tributaries were not sampled but were assumed to be equal to 20% of their suspended loads, based on the Shotover result.

Deposition in Lake Roxburgh has been monitored by cross-section survey at approximately 5-yearly intervals since 1961, when the reservoir storage volume was 101×10^6 m^3 (Webby *et al.* 1996). The earliest surveys used tagline and sounding line, whereas the latest surveys use differential GPS and echo sounder. Cores collected along the length of the lake were used by Thompson (1976) to determine the particle size of the trapped sediment and to estimate an overall bulk sediment density of 1.27 t m^{-3} via empirical relations given in Vanoni (1975). Using this bulk density, the average mass entrapment rate over the period 1961–1989 was 1.80×10^6 t per year. When combined with the trap efficiency information provided by the index sampling programme (Jowett and Hicks 1981), this indicates a robust measure of the mean annual sediment inflows to Lake Roxburgh of 2.15×10^6 t per year, which compares very favourably with the total sediment inflow of 2.36×10^6 t per year estimated independently from the tributary data. Moreover, after adjusting bedload inputs from the tributaries for abrasion (after Adams 1979) – which transforms part of the bedload to fine suspended load – a close match is achieved between the sediment inflows by size fraction (based on the tributary data) and the rate of entrapment by size fraction, at least for the sand and gravel fractions that are efficiently trapped in the reservoir (Fig. 15.10c). Such agreement by independent methods lends confidence to the overall sediment budget determination, particularly to the bedload results obtained for the Shotover River, where the agreement between sampled bedload discharges and equation predictions were not ideal. It also highlights the use of reservoirs as large-scale, long-term sampling devices.

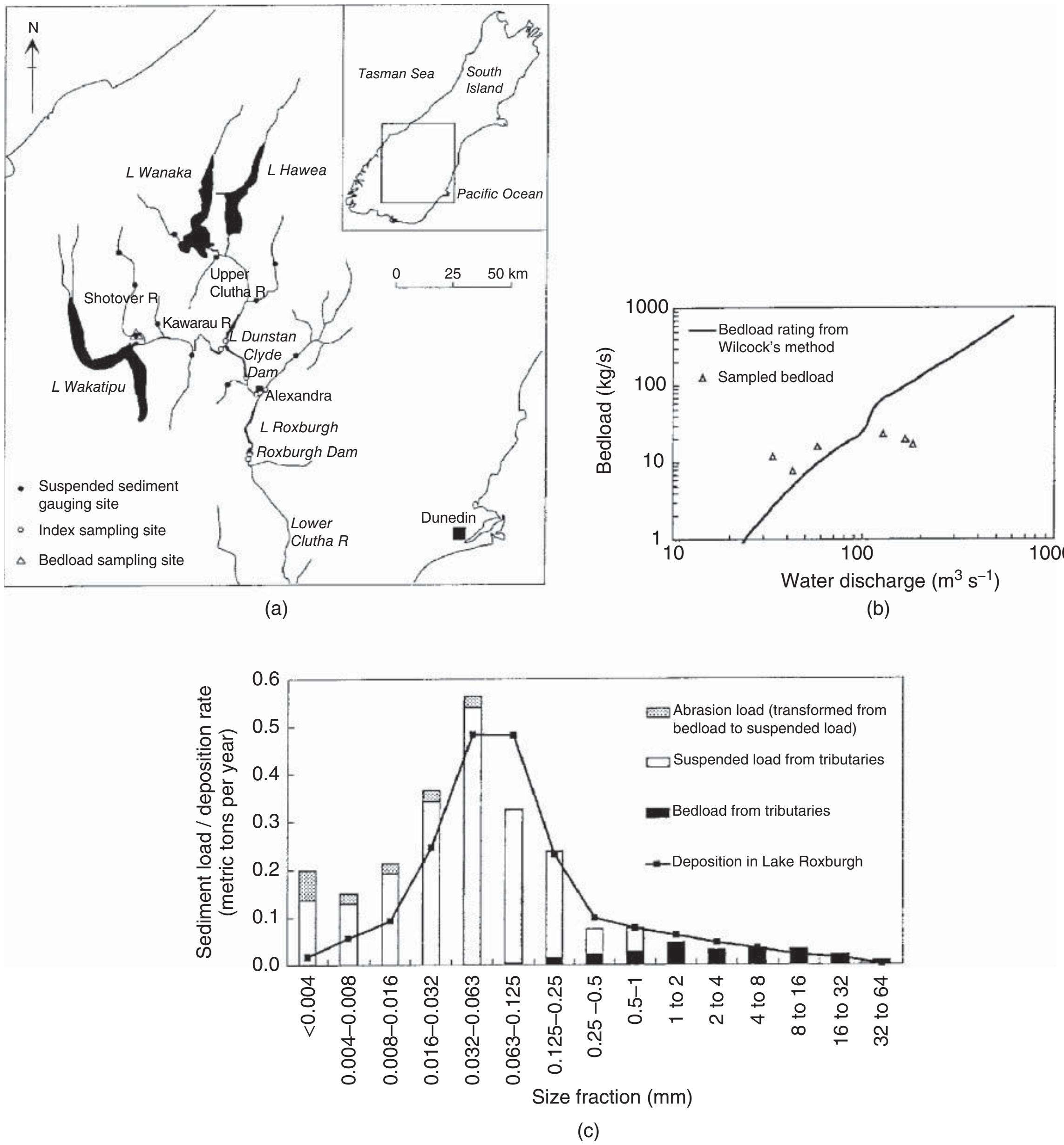


Figure 15.10 (a) Clutha River basin, South Island, New Zealand, showing hydro-dams, suspended sediment and bedload gauging sites on tributaries and index sampling sites on mainstem channels. (b) Bedload rating for Shotover River, based on Wilcock's (1977a) method, compared with bedload discharges measured with Helley–Smith sampler. (c) Annual average sediment inflows to Lake Roxburgh by size fraction, based on sediment sampling in tributaries and adjusted for abrasion (i.e. transformation of coarse bedload into suspended load), compared with average deposition rate in Lake Roxburgh by size fraction, based on reservoir surveys. See text for explanation.

Acknowledgements

We thank Mike Church, Jonathan Laronne and the volume editors for their constructive review comments. The authors' collaborative efforts were supported by the US National Science Foundation (SBR-9807195) and the New Zealand Foundation for Research Science and Technology (C09612).

References

Abbott, J.E. and Francis, J.D.R. 1977. Saltation and suspension trajectories of solid grains in a water stream. *Philosophical Transactions of the Royal Society of London* 284A: 225–254.

Abraham, D., Kuhnle, R.A. and Odgaard, A.J. 2011. Validation of bed-load transport measurements with time-sequenced bathymetric data. *Journal of Hydraulic Engineering* 137: 723–728.

Adams, J. 1979. Wear of unsound pebbles in river headwaters. *Science* 203: 171–172.

Agrawal, Y.C., McCave, I.N. and Riley J.B. 1991. Laser diffraction size analysis. In: Syvitski, J.P.M., ed., *Principles,* Methods and Application of Particle Size Analysis, Cambridge: Cambridge University Press, pp. 119–128.

Allen, T. 1990. *Particle Size Measurement*, 4th edn., London: Chapman & Hall.

Alp, M. and Ciizoglu, H.K. 2007. Suspended sediment load simulation by two artificial neural network methods using hydrometeorological data. *Environmental Modelling and Software* 22: 2–13.

American Society of Civil Engineers Task Committee. 1992. Sediment and aquatic habitat in river systems. *Journal of Hydraulic Engineering* 118: 669–687.

Andrews, E.D. 1981. Measurement and computation of bed-material discharge in a shallow sand-bed stream, Muddy Creek, Wyoming. *Water Resources Research* 17: 131–141.

Andrews, E.D. 1994. Marginal bed load transport in a gravel bed stream, Sagehen Creek, California. *Water Resources Research* 30: 2241–2250.

Andrews, E.D. and Nankervis, J.M. 1995. Effective discharge and the design of channel maintenance flows for gravel-bed rivers. *American Geophysical Union, Geophysical Monograph* 89: 151–164.

Andrews, E.D. and Smith, J.D. 1992. A theoretical model for calculating marginal bedload transport rates of gravel. In: Billi, P., Hey, R.D., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel-bed Rivers*, Chichester: John Wiley & Sons, pp. 41–48.

ASCE Task Committee. 1971. Sediment transportation mechanics: sediment discharge formulas. *Journal of the Hydraulics Division, American Society of Civil Engineers* 97(HY4): 523–567.

Ashmore, P.E. and Day, T.J. 1988. Spatial and temporal patterns of suspended-sediment yield in the Saskatchewan River Basin. *Canadian Journal of Earth Sciences* 25: 1450–1463.

Aziz, N.M. 1996. Error estimate in Einstein's suspended sediment load method. *Journal of Hydraulic Engineering* 122: 282–285.

Bagnold, R.A. 1966. *An Approach to the Sediment Transport Problem from General Physics*, US Geological Survey Professional Paper 422-I, Washington, DC: US Government Printing Office, 37pp.

Bale, A.J. and Morris, A.W. 1991. In situ size measurements of suspended particles in estuarine and coastal waters using laser diffraction. In: Syvitski, J.P.M., ed., *Principles,Methods and Application of Particle Size Analysis*, Cambridge: Cambridge University Press, pp. 197–208.

Barry, J.J., Buffington, J.M. and King, J.G. 2004. A general power equation for predicting bed load transport rates in gravel bed rivers. *Water Resources Research* 40: W10401.

Barry, J.J., Buffington, J.M., Goodwin, P., King, J.G. and Emmett, W.W. 2008. Performance of bed-load transport equations relative to geomorphic significance: predicting effective discharge and its transport rate. *Journal of Hydraulic Engineering* 134: 601–615.

Barton, J.S., Slingerland, R.L., Pittman, S. and Gabrielson, T.B. 2010. Monitoring coarse bedload transport with passive acoustic instrumentation: a field study. In: Gray, J.R., Laronne, J.B. and Marr, J.D.G., eds., *Bedload-Surrugate Monitoring Technologies*, US Geological Survey Scientific Investigations Report 2010-5091, Reston, VA: US Geological Survey, pp. 38–51, paper available online at http://pubs.usgs.gov/sir/2010/5091/papers/Barton.pdf (accessed 20 October 2015).

Bechteler, W. and Maurer M. 1991. Reliability theory applied to sediment transport formulae. In: Soulsby, R. and Bettes, R., eds., *Sand Transport in Rivers*, Rotterdam: A.A. Balkema, pp. 223–228.

Benda, L. and Dunne, T. 1997. Stochastic forcing of sediment routing and storage in channel networks. *Water Resources Research* 33: 2865–2880.

Bennett, S.J. and Best, J.L. 1995. Particle size and velocity discrimination in a sediment-laden turbulent flow using phase Doppler anemometry. *Journal of Fluids Engineering* 117: 505–511.

Beschta, R.L. 1981. Increased bag size improves Helley–Smith bedload sampler for use in streams with high sand and organic matter transport. In: Walling, D.E. and Tacconi, P., eds., *Erosion and Sediment Transport Measurement*, IAHS Publication No. 133, Wallingford: International Association of Hydrological Sciences, pp. 17–25.

Brune, R.C. 1953. Trap efficiency of reservoirs. *Transactions of the American Geophysical Union* 34(3): 407–418.

Buffington, J.M. and Montgomery, D.R. 1997. A systematic analysis of eight decades of incipient motion studies, with special reference to gravel-bedded rivers. *Water Resources Research* 33: 1993–2029.

Bunte, K. and Abt. S.R. 2005. Effect of sampling time on measured gravel bed load transport rates in a coarse-bedded stream. *Water Resources Research* 41: W11405.

Bunte, K. and Abt, S.R. 2009. *Transport Relationships Between Bedload Traps and a 3-inch Helley–Smith Sampler in Coarse Gravel-bed Streams and Development of Adjustment Functions*, Completion Report 218, Fort Collins, CO: Colorado Water Institute, Colorado State University, 140pp, http://www.cwi.colostate.edu/publications/cr/218.pdf (accessed 20 October 2015).

Bunte, K., Abt, S.R., Potyondy, J.P. and Ryan, S.E. 2004. Measurement of coarse gravel and cobble transport using a portable bedload trap. *Journal of Hydraulic Engineering* 130: 879–893.

Bunte, K., Abt, S.R., Potyondy, J.P. and Swingle, K.W. 2008. A comparison of coarse bedload transport measured with bedload traps and Helley–Smith samplers. *Geodynamica Acta* 21: 53–66.

Butcher, D.P., Ladadz, J.C., Potter, A.W.R. and White P. 1992. Reservoir sedimentation rates in the southern Pennines, United Kingdom, using reservoir sedimentation monitoring. In: Bogen, J., Walling, D.E. and Day, T., eds., *Erosion and Sediment Transport Monitoring in River Basins, Proceedings of the Oslo Symposium, 24–28 August 1992*, IAHS Publication No. 210, Wallingford: International Association of Hydrological Sciences, pp. 443–454.

Carey, W.P. and Hubbell D.W. 1986. Probability distributions for bedload transport. In: *Proceedings of the 4th Federal Inter-Agency Sedimentation Conference*, vol. 2, Washington, DC: US Government Printing Office, pp. 4.131–4.140.

Carson, M.A. and Griffiths, G.A. 1987. Bedload transport in gravel channels. *Journal of Hydrology (New Zealand)* 26: 1–151.

Carson, M.A. and Griffiths, G.A. 1989. Gravel transport in the braided Waimakariri River: mechanisms, measurements and predictions. *Journal of Hydrology* 109: 201–220.

Childers, D. 1991. Sampling differences between the Helley–Smith and BL-84 bedload samplers. In: *Proceedings of the 5th Federal Inter-Agency Sedimentation Conference, Las Vegas*, pp. 6.31–6.38.

Childers, D. 1999. *Field Comparison of Six Pressure-difference Bedload Samplers in High Energy Flow*, US Geological Survey Water Resources Investigations Report 92-4068, Vancouver, WA: US Geological Survey, http://pubs.usgs.gov/wri/1992/4068/report.pdf (accessed 20 October 2015).

Childers, D., Huanjing, G. and Gangyan, Z. 1989. Field comparison of bedload samplers from the United States of America and the Peoples Republic of China. In: *Proceedings of the 4th International Symposium on River Sedimentation*, vol. 2, Beijing: China Ocean Press, pp. 1309–1316.

Christian, R. and Thompson, S.M. 1978. Loop rating and grading of suspended sediment in the Mararoa. *Journal of Hydrology (New Zealand)* 17: 50–53.

Church, M.A. and Hassan, M.A. 2005. Upland gravel-bed rivers with low sediment transport. In: Garcia, C. and Batalla, R., eds., *Catchment Dynamics and River Processes*, Amsterdam: Elsevier, pp. 141–168.

Church, M.A., Miles, M. and Rood, K. 1986. *Sediment Transfer Along Mackenzie River: a Feasibility Study*, Sediment Survey Section Report IWD-WNR(NWT)-WRB-SS-871, Ottawa: Environment Canada, Inland Waters Directorate, Western and Northern Region.

Church, M.A., Wolcott, J.F. and Fletcher, W.K. 1991. A test of equal mobility in fluvial sediment transport: behavior of the sand fraction. *Water Resources Research* 27: 2941–2951.

Church, M.A., Ham, E., Hassan, M. and Slaymaker, O. 1999. Fluvial clastic sediment yield in Canada: scaled analysis. *Canadian Journal of Earth Sciences* 36: 1267–1280.

Cioffi, F. and Gallerano F. 1991. Velocity and concentration profiles of solid particles in a channel with movable and erodible bed. *Journal of Hydraulic Research* 29: 387–401.

Cleveland, W.S. 1979. Robust locally weighted regression and smoothing scatterplots. *Journal of the American Statistical Association* 74: 829–836.

Coakley, J.P. and Syvitski, J.P.M. 1991. SediGraph technique. In: Syvitski, J.P.M. ed., *Principles,* Methods and Application of Particle Size Analysis, Cambridge: Cambridge University Press, pp. 129–142.

Cohn, T.A. 1995. Recent advances in statistical methods for the estimation of sediment and nutrient transport in rivers. *Reviews of Geophysics* 33(Suppl. 2): 1117–1123.

Cohn, T.A., DeLong, L.L., Gilroy, E.J., Hirsch R.M. and Wells, D.K. 1989. Estimating constituent loads. *Water Resources Research* 25: 937–942.

Colby, B.R. and Hembree, B.H. 1955. *Computations of Total Sediment Discharge, Niobrara River Near Cody, Nebraska*, US Geological Survey Water-Supply Paper 1357, Washington, DC: US Government Printing Office, 187pp.

Costa, J.E. 1988. Rheologic, geomorphic and sedimentologic differentiation of water floods, hyperconcentrated flows and debris flow. In: Baker, V.R., Kochel, R.C. and Patton P.C., eds., *Flood Geomorphology*, Chichester: John Wiley & Sons, pp. 113–122.

Crawford, C.G. 1991. Estimation of suspended-sediment rating curves and mean suspended-sediment loads. *Journal of Hydrology* 129: 331–348.

Csoma, J. 1973. Reliability of bed-load sampling. *Proceedings of the International Association for Hydraulics Research* 2: 97–107.

Cui, Y. 2007. The unified gravel-sand (TUGS) model: simulating sediment transport and gravel/sand grain size distributions in gravel-bedded rivers. *Water Resources Research* 43: W10436.

Cui, Y. and Parker, G. 2005. Numerical model of sediment pulses and sediment supply disturbances in mountain rivers. *Journal of Hydraulic Engineering* 131: 646–656.

Cui, Y., Parker, G., Lisle, T.E., Pizzuto, J.E. and Dodd, A.M. 2005. More on the evolution of bed material waves in alluvial rivers. *Earth Surface Processes and Landforms* 30: 107–114.

Czuba, J.A., Best, J.L., Oberg, K.A., Parsons, D.R., Jackson, P.R., Garcia, M.C. and Ashmore, P. 2011. Bed morphology, flow structure and sediment transport at the outlet of Lake Huron and in the upper St. Clair River. *Journal of Great Lakes Research* 37: 480–493.

Davies, A.P. 1900. *Hydrography of Nicaragua*, In: US Geological Survey 20th Annual Report 1898–99, Washington, DC: US Government Printing Office, pp. 563–637.

Day, T.J. 1988. Evaluation of long term suspended sediment records for selected Canadian rivers. In: *Sediment Budgets, Proceedings of the Porto Alegre Symposium, December 1988*, IAHS Publication No. 174, Wallingford: International Association of Hydrological Sciences, pp. 101–107.

De Lange, W.P., Healy T.R. and Darlan, Y. 1997. Reproducibility of sieve and settling tube textural determinations for sand-sized beach sediment. *Journal of Coastal Research* 13: 73–80.

De Vries, M. 1973. *On Measuring Discharge and Sediment Transport*, Delft Hydraulics Laboratory Publication 106, Delft: Delft Hydraulics Laboratory.

Dietrich, W.E. and Whiting P.J. 1989. Boundary shear stress and sediment transport in river meanders of sand and gravel. In: Ikeda, S. and Parker, G., eds., *River Meandering*, Washington, DC: American Geophysical Union, pp. 1–50.

Dobson, G.C. and Johnson, J.W. 1940. Studying sediment loads in natural streams. *Civil Engineering* 10: 93–96.

Druffel, L., Emmett, W.W., Schneider, V.R. and Skinner J.V. 1976. *Laboratory Hydraulic Calibration of the Helley–Smith Bedload Sediment Sampler*, US Geological Survey Open-File Report 76-752, Bay St. Louis, MS: US Geological Survey.

Duan, N. 1983. Smearing estimate: a non-parametric retransformation method. *Journal of the American Statistical Association* 78: 605–610.

Du Boys, P.F.D. 1879. Le Rhône et les rivières à lit affouillable. *Annales des Ponts et Chaussées, Série* 5, 18: 141–195.

Edwards, T.K. and Glysson, G.D. 1999. *Techniques of Water-Resources Investigations of the US Geological Survey, Book 3, Applications of Hydraulics, Chapter C2, Field Methods for Measurement of Fluvial Sediment*, TWRI 3-C2, Reston, VA: US Geological Survey, 89pp.

Einstein, H.A. 1937. Die Eichung des im Rhein verwendeten Geschiebefängers [Calibrating the bedload trap used in the Rhine]. *Schweizerische Bauzeitung* 110(12): 29–32.

Einstein, H.A. 1944. *Bedload Transportation in Mountain Creek*, Technical Paper SCS-TP-55, Washington, DC: US Department of Agriculture, Soil Conservation Service, 54 pp.

Einstein, H.A. 1948. Determination of rates of bedload movement. In: *Proceedings of the 1st Federal Inter-Agency Sedimentation Conference*, pp. 75–93.

Einstein, H.A. 1950. *The Bedload Function for Sediment Transportation in Open Channels*, Technical Bulletin 1026, Washington, DC: US Department of Agriculture, Soil Conservation Service, 78pp.

Einstein, H.A., Anderson, A.G. and Johnson, J.W. 1940. A distinction between bed load and suspended load. *Transactions of the American Geophysical Union* 21: 628–632.

Emmett, W.W. 1980. *A Field Calibration of the Sediment-trapping Characteristics of the Helley–Smith Bedload Sampler*, US Geological Survey Professional Paper 1139, Washington, DC: US Government Printing Office.

Emmett, W.W. and Wolman, M.G. 2001. Effective discharge and gravel-bed rivers. *Earth Surface Processes and Landforms* 26: 1369–1380.

Engelund, F. and Hansen, E. 1967. *A Monograph on Sediment Transport in Alluvial Streams*, Copenhagen: Teknisk Vorlag, 62pp.

Ferguson, R.I. 1986. River loads underestimated by rating curves. *Water Resources Research* 22: 74–76.

Ferguson, R.I. 1987. Accuracy and precision of methods for estimating river loads. *Earth Surface Processes and Landforms* 12: 95–104.

Ferguson, R.I. and Ashworth, P.J. 1992. Spatial patterns of bedload transport and channel change in braided and near-braided rivers. In: Billi, P., Hey R.D., Thorne C.R. and Tacconi, P., eds. *Dynamics of Gravel-bed Rivers*, Chichester: John Wiley & Sons, pp. 477–496.

Ferguson, R. and Church, M. 2009. A critical perspective on 1-D modeling of river processes: gravel load and aggradation in lower Fraser River. *Water Resources Research* 45: W11424.

Ferguson, R.I. and Paola C. 1997. Bias and precision of percentiles of bulk grain size distributions. *Earth Surface Processes and Landforms* 22: 1061–1077.

Ferguson, R.I., Church, M. and Weatherly, H. 2001. Fluvial aggradation in Vedder River: testing a one-dimensional sedimentation model. *Water Resources Research* 37: 3331–3347.

Fienberg, K., Singh, A., Foufoula-Geeorgiou, E., Jerolmack, D. and Marr, J.D.G. 2010. A theoretical framework for interpreting and quantifying the sampling time dependence of gravel bedload transport rates. In: Gray, J.R., Laronne, J.B. and Marr, J.D.G., eds., *Bedload-Surrugate Monitoring Technologies*, US Geological Survey Scientific Investigations Report 2010-5091, Reston, VA: US Geological Survey, pp. 171–184, paper available online at http://pubs.usgs.gov/sir/2010/5091/papers/Fienberg.pdf (accessed 20 October 2015).

Foster, I.D.L., Millington R. and Grew, R.G. 1992. The impact of particle size controls on stream turbidity measurements; some implications for suspended sediment yield estimation. In: Bogen, J., Walling, D.E. and Day T.J., eds., *Erosion and Sediment Transport Monitoring Programmes in River Basins, Proceedings of the Oslo Symposium, 24–28 August 1992*, IAHS Publication No. 210, Wallingford: International Association of Hydrological Sciences, pp. 51–62.

Frey, P. and Church. M. 2011. Bedload: a granular phenomenon. *Earth Surface Processes and Landforms* 36: 58–69.

Fuller, I.C., Large, A.R.G., Charlton, M.E., Heritage, G.L., Milan, D.J. 2003. Reach-scale sediment transfers: an evaluation of two morphological budgeting approaches. *Earth Surface Processes and Landforms* 28: 889–903.

Gabel, S.L. 1993. Geometry and kinematics of dunes during steady and unsteady flows in the Calamus River, Nebraska, USA. *Sedimentology* 40: 237–269.

Gaeuman, D. and Jacobson, R.B. 2007. Field assessment of alternative bed-load transport estimators. *Journal of Hydraulic Engineering* 133: 1319–1328.

Gaeuman, D., Andrews, E.D., Krause, A. and Smith, W. 2009. Predicting fractional bed load transport rates: application of the Wilcock–Crowe equations to a regulated gravel bed river. *Water Resources Research* 45: 1–15.

Galai Production. 1997. *WCIS-100 Particle Size Analyzer Operation Manual*, Migdal Haemek: Galai Production.

Ganti, V., Meerschaert, M.M., Foufoula-Georgiou, E., Viparelli, E. and Parker, G. 2010. Normal and anomalous diffusion of gravel tracer particles in rivers. *Journal of Geophysical Research* 115: F00A12.

Garcia, C., Laronne, J.B. and Sala, M. 1999. Variable source areas of bedload in a gravel-bed stream. *Journal of Sedimentary Research* 69: 27–31.

Gartner, J.W., 2003. Estimation of suspended solids concentrations based on acoustic backscatter intensity: theoretical background. In: Gray, J.R. and Glysson, G.D., eds., *Proceedings of the Federal Interagency Workshop on Turbidity and Other Sediment Surrogates, April 30–May 2, 2002, Reno, Nevada*, US Geological Survey Circular 1250, Reston, VA: US Geological Survey, 136pp.

Gartner, J.W. and Cheng, R.T. 2001. The promises and pitfalls of estimating total suspended solids based on backscatter intensity from acoustic Doppler current profilers. In: *Proceedings of the 7th Federal Inter-Agency Sedimentation Conference*, pp. III-119–III-126.

Gaudet, J.M., Roy, A.G. and Best, J.L. 1994. Effect of orientation and size of Helley–Smith sampler on its efficiency. *Journal of Hydraulic Engineering* 120: 758–766.

Gaweesh, M.T.K. and Van Rijn, L.C. 1994. Bed-load sampling in sand-bed rivers. *Journal of Hydraulic Engineering* 120: 1364–1384.

Gentien, P., Lunven, M., Lehaitre, M. and Duvent J.L. 1995. In situ depth profiling of particle sizes. *Deep-Sea Research* 42: 1297–1312.

Gilbert, G.K. 1914. *The Transportation of Debris by Running Water*, US Geological Survey Professional Paper 86, Washington, DC: US Government Printing Office.

Gill, M.A. 1988. Planning the useful life of a reservoir. *Water Power and Dam Construction* (May): 46–47.

Gippel, C.J. 1989. The use of turbidimeters in suspended sediment research. *Hydrobiologia* 176/177: 465–480.

Gippel, C.J. 1995. Potential of turbidity monitoring for measuring the transport of suspended solids in streams. *Hydrological Processes* 9: 83–97.

Gomez, B. 1983. Temporal variations in bedload transport rates: the effect of progressive bed armouring. *Earth Surface Processes and Landforms* 8: 41–54.

Gomez, B. 1991. Bedload transport. *Earth Science Reviews* 31: 89–132.

Gomez, B. 1995. Bedload transport and changing grain size distributions. In: Gurnell, A.M. and Petts, G., eds., *Changing River Channels*, Chichester: John Wiley & Sons, pp. 177–199.

Gomez, B. 2006. The potential rate of bed-load transport. *Proceedings of the National Academy of Sciences of the United States of America* 103: 17170–17173.

Gomez, B. and Church, M. 1989. An assessment of bedload sediment transport formulae for gravel-bed rivers. *Water Resources Research* 25: 1161–1186.

Gomez, B. and. Troutman, B.M. 1997. Evaluation of process errors in bed load sampling using a dune model. *Water Resources Research* 33: 2387–2398.

Gomez, B., Naff, R.L. and Hubbell, D.W. 1989. Temporal variations in bedload transport rates associated with the migration of bedforms. *Earth Surface Processes and Landforms* 14: 135–156.

Gomez, B., Hubbell, D.W. and Stevens, H.H. 1990. At-a-point bedload sampling in the presence of dunes. *Water Resources Research* 26: 2717–2731.

Gomez, B., Emmett, W.W. and Hubbell, D.W. 1991. Comments on bedload sampling in small rivers. In: *Proceedings of the 5th*

Federal Inter-Agency Sedimentation Conference, Las Vegas, vol. 1, pp. 2.65–2.72.

Gomez, B., Mertes, L.A.K., Phillips, J.D., Magilligan, F.J. and James, L.A. 1995. Sediment characteristics of an extreme flood: 1993 upper Mississippi River valley. *Geology* 23: 963–966.

Gomez, B., Cui, Y., Kettner, A.J., Peacock, D.H. and Syvitski, J.P.M. 2009. Simulating changes to the sediment transport regime of the Waipaoa River driven by climate change in the twenty-first century. *Global and Planetary Change* 67: 153–166.

Gray, J.R. and Fisk G.G. 1992. *Monitoring Radionuclide and Suspended-sediment Transport in the Little Colorado River Basin, Arizona and New Mexico, USA*, IAHS Publication No. 210, Wallingford: International Association of Hydrological Sciences, pp. 505–516.

Gray, J.R. and Gartner, J.W. 2009. Technological advances in suspended sediment surrogate monitoring. *Water Resources Research* 45: W00D29.

Gray, J.R. and Gartner, J.W. 2010. Surrogate technologies for monitoring suspended-sediment transport in rivers. In: Poleto, C. and Charlesworth, S., eds., *Sedimentology of Aqueous Systems*, Oxford: Blackwell, pp. 3–45.

Gray, J.R., Laronne, J.B. and Marr J.D.G., eds. 2010. *Bedload-Surrogate Monitoring Technologies*, US Geological Survey Scientific Investigations Report 2010-5091, Reston, VA: US Geological Survey, http://pubs.usgs.gov/sir/2010/5091 (accessed 20 October 2015).

Green, M.O., Dolphin, T.J., Swales, A. and Vincent, C.E. 1999. Transport of mixed-size sediments in a tidal channel. In: Kraus, N.C. and McDougal, W.G., eds., *Coastal Sediments '99: Proceedings of the 4th International Symposium on Coastal Engineering and Science of Coastal Sediment Processes, Hauppauge, Long Island, New York, June 21-23, 1999*, vol. 1, Reston, VA: American Society of Civil Engineers, pp. 644–658.

Guy, H.P. 1969. *Techniques of Water-Resources Investigations of the US Geological Survey, Book 5, Laboratory Analyses, Chapter C1, Laboratory Theory and Methods for Sediment Analysis*, TWRI 5-C1, Reston, VA: US Geological Survey, 58pp.

Haan, C.T. 1977. *Statistical Methods in Hydrology*, Ames, IA: Iowa State University Press, 378pp.

Habersack, H.M., Nachtnebel, H.-P. and Laronne, J.B. 1998. Hydraulic efficiency of a slot sampler: flow velocity measurements in the Drau River, Austria. In: Klingeman, P.C., Beschta, R.L., Komar, P.D. and Bradley, J.B, eds., *Gravel-bed Rivers in the Environment*, Highlands Ranch, CO: Water Resources Publications, pp. 749–754.

Hansen, E.A. 1973. In-channel sedimentation basins, a possible tool in trout habitat management. *Progressive Fish Culturist* 35: 136–142.

Haschenburger, J.K. and Church, M.A. 1998. Bed material transport estimated from the virtual velocity of sediment. *Earth Surface Processes and Landforms* 23: 791–808.

Haschenburger, J.K. and Wilcock, P.R. 2003. Partial transport in a natural gravel-bed channel. *Water Resources Research* 39(1): 1020.

Hassan, M.A and Church, M.A. 2000. Experiments on surface structure and partial sediment transport on a gravel bed. *Water Resources Research* 36: 1885–1895.

Hassan, M.A. and Church, M.A. 2001. Sensitivity of bedload transport in Harris Creek: seasonal and spatial variation over a cobble-gravel bar. *Water Resources Research* 37: 813–825.

Hayward, J.A. 1980. *Hydrology and Stream Sediment from Torlesse Stream Catchment*, Special Publication 17, Canterbury, NZ: Tussock Grasslands and Mountain Lands Institute, Lincoln College, 236pp.

Hayward, J.A. and Sutherland, A.J. 1974. The Torlesse Stream vortex-tube sediment trap. *Journal of Hydrology (New Zealand)* 13: 41–53.

Helley, E.J. and Smith, W. 1971. *Development and Calibration of a Pressure-difference Bedload Sampler*, US Geological Survey Open-File Report 73-108, Menlo Park, CA: US Geological Survey, pp. 21.

Hicks, D.M. 1994. *Landuse Effects on Magnitude–Frequency Characteristics of Storm Sediment Yields: Some New Zealand Examples*, IAHS Publication No. 224, Wallingford: International Association of Hydrological Sciences, pp. 395–402.

Hicks, D.M. and Duncan, M.J. 1997. The efficiency of depth-integrating samplers in sampling the suspended sand load in gravel bed rivers. *Journal of Hydrology* 201: 138–160.

Hicks, D.M., Gomez, B. and Trustrum, N.A. 2000. Erosion thresholds and suspended sediment yields, Waipaoa River basin, New Zealand. *Water Resources Research* 36: 1129–1142.

Hoitink, A.J.F. and Hoekstra, P. 2005. Observations of suspended sediment from ADCP and OBS measurements in a mud-dominated environment. *Coastal Engineering* 52: 103–118.

Holmes, R.R. 2010. Measurement of bedload transport in sand-bed rivers: a look at two indirect sampling methods. In: Gray, J.R., Laronne, J.B. and Marr, J.D.G., eds., *Bedload-Surrugate Monitoring Technologies*, US Geological Survey Scientific Investigations Report 2010-5091, Reston, VA: US Geological Survey, pp. 236–252, paper available online at http://pubs.usgs.gov/sir/2010/5091/papers/Holmes.pdf (accessed 20 October 2015).

Horowitz, A.J. 1985. *A Primer on Trace Metal Sediment Chemistry*, US Geological Survey Water-Supply Paper 2277, Washington, DC: US Government Printing Office.

Hubbell, D.W. 1964. *Apparatus and Techniques for Measuring Bedload*, US Geological Survey Water-Supply Paper 1748, Washington, DC: US Government Printing Office.

Hubbell, D.W. 1987. Bedload sampling and analysis. In: Thorne, C.R., Bathurst, J.C. and Hey, R.D., eds., *Sediment Transport in Gravel-bed Rivers*, Chichester: John Wiley & Sons, pp. 89–120.

Hubbell, D.W. and Matejka, D.Q. 1959. *Investigations of Sediment Transportation, Middle Loup River at Dunning, Nebraska, with Application of Data from Turbulence Flume*, US Geological Survey Water-Supply Paper 1476, Washington, DC: US Government Printing Office.

Hubbell, D.W., Stevens, H.H., Skinner, J.V. and Beverage, J.P. 1981. Recent refinements in calibrating bedload samplers. In: *Water Forum '81*, New York: American Society of Civil Engineers, pp. 128–140.

Hubbell, D.W., Stevens, H.H., Skinner, J.V. and Beverage, J.P. 1985. New approach to calibrating bedload samplers. *Journal of Hydraulic Engineering* 111: 677–694.

Humphreys, A.A. and Abbot H.L. 1861. Report Upon the Physics and Hydraulics of the Mississippi River; Upon the Protection of the Alluvial Region Against Overflow; and Upon the Deepening of the Mouths Based Upon Surveys and Investigations, Professional Papers of the Corps of Topographical Engineers, US Army, No. 4, Washington, DC: US Government Printing Office, 456pp.

Jackson, W.L. and Beschta, R.L. 1982. A model of two-phase bedload transport in an Oregon Coast Range stream. *Earth Surface Processes and Landforms* 7: 517–527.

James, A.L. 2010. Secular sediment waves, channel bed waves and legacy sediment. *Geography Compass* 4(6): 576–598.

Johnson, C.W., Engleman, R.L., Smith, J.P. and Hanson, C.L. 1977. Helley–Smith bedload samplers. *Journal of the Hydraulics Division, American Society of Civil Engineers* 103: 1217–1221.

Jowett, I.G. and Hicks, D.M. 1981. Surface, suspended and bedload sediment – Clutha River system. *Journal of Hydrology (New Zealand)* 20: 121–130.

Kelsey, H.M. 1980. A sediment budget and an analysis of geomorphic processes in the Van Duzen River Basin, northern coastal California 1941–1975. *Geological Society of America Bulletin, Part II* 91: 1119–1216.

Kisi, O. 2005. Suspended sediment estimation using neuro-fuzzy and neural network approaches. *Hydrological Sciences Journal* 50: 683–696.

Kisi, O., Karahan, M.E. and Sen, Z. 2006. River suspended sediment modeling using a fuzzy logic approach. *Hydrological Processes* 20: 4351–4362.

Kondolf, G.M. and Wilcock, P.R. 1996. The flushing flow problem: defining and evaluating objectives. *Water Resources Research* 32: 2589–2599.

Krumbein, W.C. and Pettijohn F.J. 1938. *Manual of Sedimentary Petrography*, New York: Appleton-Century.

Kuhnle, R.A. 1996. Unsteady transport of sand and gravel mixtures. In: Carling, P.A. and Dawson, M., eds., *Advances in Fluvial Dynamics and Stratigraphy*, Hoboken, NJ: John Wiley & Sons, pp. 183–201.

Kuhnle, R.A. 1998. Statistics of sediment transport in Goodwin Creek. *Journal of Hydraulic Research* 124: 1109–1114.

Lane, E.W. and Koelzer, V.A. 1943. *A Study of Methods Used in Measurement and Analysis of Sediment Loads in Streams. Report No. 9: Density of Sediments Deposited in Reservoirs*, Iowa City, IA: St. Paul US Engineer District Sub-Office, Hydraulic Laboratory, University of Iowa.

Lane, S.N., Richards, K.S. and Chandler, J.H. 1995. Morphological estimation of the time-integrated bed load transport rate. *Water Resources Research* 31: 761–772.

Lane, S.N., Westaway, R.M. and Hicks, D.M. 2003. Estimation of erosion and deposition volumes in a large, gravel-bed, braided river using synoptic remote sensing. *Earth Surface Processes and Landforms* 28: 249–271.

Laronne, J.B. 1990. Probability distribution of event sediment yields in the Northern Negev, Israel. In: Boardman, J., Foster, I. and Dearing, J., eds., *Soil Erosion in Agricultural Land*, Chichester: John Wiley & Sons, pp. 481–492.

Laronne, J.B. and Wilhelm, R. 2001. Shifting stage–volume curves: predicting event sedimentation rate based on reservoir stratigraphy. In: Anthony, D., Ethridge F., Harvey, M., Laronne J.B. and Mosley M.P., eds., *Applying Geomorphology to Environmental Management*, Highlands Ranch, CO: Water Resources Publications, pp. 33–54.

Lawler, D.M. and Brown, R.M. 1992. A simple and inexpensive turbidity meter for the estimation of suspended sediment concentrations. *Hydrological Processes* 6: 159–168.

Lenzi, M.A., D'Agostino, V. and Billi, P. 1999. Bedload transport in the instrumented catchment of the Rio Cordon. Part I: Analysis of bedload records, conditions and threshold of bedload entrainment. *Catena* 36: 171–190.

Leopold, L.B. and Emmett, W.W. 1997. *Bedload and River Hydraulics Inferences from the East Fork River, Wyoming*, US Geological Survey Professional Paper 1583, Denver, CO: US Geological Survey, 52pp.

Lewis, J. 1996. Turbidity-controlled suspended sediment sampling for runoff-event load estimation. *Water Resources Research* 32: 2299–2310.

Lisle, T.E. 1995. Particle size variations between bed load and bed material in natural gravel bed channels. *Water Resources Research* 31: 1107–1118.

Lisle, T.E. and Madej, M.A. 1992. Spatial variation in armouring in a channel with high sediment supply. In: Billi, P., Hey, R.D., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel-bed Rivers*, Chichester: John Wiley & Sons, pp. 277–296.

Lisle, T.E., Nelson, J.E., Pitlick, J., Madej, M.A. and Barkett, B.L. 2000. Variability of bed mobility in natural gravel-bed channels and adjustments to sediment load at local and reach scales. *Water Resources Research* 36: 3743–3755.

Mahmood, K. 1980. *Verification of Sediment Transport Functions – Missouri River*, US Army Corps of Engineers, Missouri River Division Sediment Series Report 19.

Malhotra, V.M. 1967. The Alpine airjet sieve: a new method of determining the fineness of cement. *Indian Concrete Journal* 41: 305–309.

Maneux, E., Probst, J.L., Veyssey, E. and Etcheber, H. 2001. Assessment of dam trapping efficiency from water residence time: application to fluvial sediment transport in the Adour, Dordogne and Garonne River basins (France). *Water Resources Research* 37: 801–811.

Mao, L., Comiti, F. and Lenzi, M.A. 2010. Bedload dynamics in steep mountain rivers: insights from the Rio Cordon Experimental Station (Italian Alps). In: Gray, J.R., Laronne, J.B. and Marr, J.D.G., eds., *Bedload-Surrugate Monitoring Technologies*, US Geological Survey Scientific Investigations Report 2010-5091, Reston, VA: US Geological Survey, pp. 253–265, paper available online at http://pubs.usgs.gov/sir/2010/5091/papers/Mao.pdf (accessed 20 October 2015).

Martin, Y. 2003. Evaluation of bed load transport formulae using field evidence from the Vedder River, British Columbia. *Geomorphology* 53: 75–95.

Martin, Y. and Church, M.A. 1995. Bed-material transport estimated from channel surveys: Vedder River, British Columbia. *Earth Surface Processes and Landforms* 20: 347–361.

McCave, I.N. and Syvitski, J.P.M. 1991. Principles and methods of geological particle size analysis. In: Syvitski, J.P.M., ed., *Principles,* Methods and Application of Particle Size Analysis, Cambridge: Cambridge University Press, pp. 1–21.

McHenry, J.R. 1971. Discussion of sediment measurement techniques: B. Reservoir deposits. *Journal of the Hydraulics Division, American Society of Civil Engineers* 97: 1253–1257.

McLean, D.G. and Church, M.A. 1999. Sediment transport along lower Fraser River 2. Estimates based on the long term gravel budget. *Water Resources Research* 35: 2549–2559.

McLean, D.G., Church, M.A. and Tassone, B. 1999. Sediment transport along lower Fraser River 1. Measurement and hydraulic computations. *Water Resources Research* 35: 2533–2548.

Meade, R.H., Yuzyk, T.R. and Day, T.J. 1990. Movement and storage of sediment in rivers of the United States and Canada. In: Wolman, M.G.

and Riggs, H.C., eds., *Surface Water Hydrology*, Boulder, CO: Geological Society of America, pp. 255–280.

Mertes, L.A.K., Smith, M.O. and Adams, J.B. 1993. Estimating suspended sediment concentrations in surface waters of the Amazon River wetlands from Landsat images. *Remote Sensing of the Environment* 43: 281–301.

Mianaei, S.J. and Keshavarzi, A.R. 2010. Prediction of riverine suspended sediment discharge using fuzzy logic algorithms and some implications for estuarine settings. *Geo-Marine Letters* 30: 30–35.

Milhous, R.T. 1973. *Sediment Transport in a Gravel-bottomed Stream*, unpublished PhD Thesis, Oregon State University, Corvallis, OR, 232pp.

Miller, C.R. 1951. *Analysis of Flow-duration, Sediment-rating Curve Method of Computing Sediment Yield*, US Bureau of Reclamation Technical Report, Denver, CO: US Bureau of Reclamation, 55pp.

Miller, C.R. 1953. *Determination of Unit Weight of Sediment for Use in Sediment Volume Computations*, Memorandum, Denver, CO: US Bureau of Reclamation.

Mohrig, D. and Smith, J.D. 1996. Predicting the migration rate of subaqueous dunes. *Water Resources Research* 32: 3207–3217.

Moll, J.R., Schilperoort, T. and Leeuw, A.J. 1987. Stochastic analysis of bedform dimensions. *Journal of Hydraulic Research* 25: 465–479.

Moog, D.B. and Whiting, P.J. 1998. Annual hysteresis in bed load rating curves. *Water Resources Research* 34: 2393–2399.

Morris, L. and Fan, J. 1998. *Reservoir Sedimentation Handbook: Design and Management of Dams, Reservoirs and Watersheds for Sustainable Use*, New York: McGraw-Hill, 704pp.

Mount, N.J. and Abrahart, R.J. 2011. Load or concentration, logged or unlogged? Addressing ten years of uncertainty in neural network suspended sediment prediction. *Hydrological Processes* 25: 3144–3157.

Mülhofer, L. 1933. Untersuchungen über die Schwebstoff- und Geschiebeführung des Inn nächst Kirchbichl, Tirol [Investigations into suspended load and bedload of the River Inn, near Kirchbichl, Tirol], *Die Wasserwirtschaft* 1–6.

Müller, R. 1937. Überprufung des Geschiebegesetzes und der Berechnungsmethode der Versuchsanstalt für Wasserbau an der E.T.H. mit Hilfe der direkten Geschiebemessungen am Rhein [Verification of the bedload law and calculation method of the Hydraulic Experiment Station E.T.H. by direct bedload measurements on the Rhine], *Schweizerische Bauzeitung* 110(12–15): 33–36.

Nanson, G.C. 1974. Bedload and suspended-load transport in a small, steep mountain stream. *American Journal of Science* 274: 471–486.

Neff, E.L. 1967. *Discharge Frequency Compared to Long-term Sediment Yields*, IAHS Publication No. 75, Wallingford: International Association of Hydrological Sciences, pp. 236–242.

Neill, C.R. 1971. River bed transport related to meander migration rates. *Journal of the Waterways, Harbors and Coastal Engineering Division, American Society of Civil Engineers* 97: 783–786.

Neill, C.R. 1987. Sediment balance considerations linking long-term transport and channel processes. In: Thorne, C.R., Bathurst, J.C. and Hey, R.D., eds., *Sediment Transport in Gravel-bed Rivers,* Chichester: John Wiley & Sons, pp. 225–240.

Nelson, M.E. and Benedict, P.C. 1950. Measurement and analysis of suspended-sediment loads in streams. *Transactions of the American Society of Civil Engineers* 116: 891–918.

Nesper, F. 1937. Ergebnisse der Messungen über die Geschiebe- und Schlammführung des Rheins an der Brugger Rheinbrücke [Results of bedload and silt movement observations on the Rhine at the Brugg Bridge], *Schweizerische Bauzeitung* 110(12–15).

Nevin, C. 1946. Competency of moving water to transport debris. *Geological Society of America Bulletin* 57: 651–674.

Newson, M. 1980. The geomorphological effectiveness of floods: a contribution stimulated by two recent events in mid-Wales. *Earth Surface Processes and Landforms* 5: 1–16.

Nikora, V.I. and Hicks, D.M. 1997. Scaling relations for sandwave development in unidirectional flow. *Journal of Hydraulic Engineering* 123: 1152–1156.

Nikora, V., Habersack, H.M., Huber, T. and McEwan, I. 2002. On bed particle diffusion in gravel-bed flows under weak bed-load transport. *Water Resources Research* 38(6): 17-1–17-9.

Nittrouer, J.A., Allison, M.A. and Campanella, R. 2008. Bedform transport rates for the lowermost Mississippi River. *Journal of Geophysical Research* 113: F03004.

Olive, L.J. and Rieger, W.A. 1988. *An Examination of the Role of Sampling Strategies in the Study of Suspended Sediment Transport*, IAHS Publication No. 174, Wallingford: International Association of Hydrological Sciences, pp. 259–267.

Ongley, E.D., Bynoe, M.C. and Percival, J.B. 1982. Physical and geochemical characteristics of suspended solids, Wilton Creek, Ontario. *Hydrobiologia* 91: 41–57.

Ongley, E.D., Yuzyk, T.R. and Krishnappan, B.G. 1990. Vertical and lateral distribution of fine-grained particulates in prairie and cordilleran rivers: sampling implications for water quality programs. *Water Research* 24: 303–312.

Parker, G. 1990. Surface-based bedload transport relation for gravel rivers. *Journal of Hydraulic Research* 28: 417–436.

Parker, G., Klingerman, P.C. and McLean, D.G. 1982. Bedload and size distribution in paved gravel-bed streams. *Journal of the Hydraulics Division, American Society of Civil Engineers* 108: 544–571.

Parker, R.S. and Troutman, B.M. 1989. Frequency distribution for suspended sediment loads. *Water Resources Research* 25: 1567–1574.

Parshall, R.L. 1952. Model and prototype studies of sand traps. *Transactions of the American Society of Civil Engineers* 117: 204–212.

Phillips, J.M. and Walling, D.E. 1995a. Measurement in situ of the effective particle-size characteristics of fluvial suspended sediment by means of a field-portable laser backscatter probe: some preliminary results. *Marine and Freshwater Research* 46: 349–357.

Phillips, J.M. and Walling, D.E. 1995b. An assessment of the effects of sample collection, storage and resuspension on the representativeness of measurements of the effective particle size distribution of fluvial suspended sediment. *Water Research* 29: 2498–2508.

Pitlick, J. 1988. Variability of bed load measurements. *Water Resources Research* 24: 173–177.

Pitlick, J., Cui, Y. and Wilcock, P. 2009. *Manual for Computing Bed Load Transport Using BAGS (Bedload Assessment for Gravel-bed Streams) Software*, General Technical Report RMRS-GTR-223. Fort Collins, CO: US Department of Agriculture, Forest Service, Rocky Mountain Research Station, 45pp., http://www.fs.fed.us/rm/pubs/rmrs_gtr223.pdf (accessed 20 October 2015).

Poreh, M., Sagiv, A. and Seginer, I. 1970. Sampling efficiency of slots. *Journal of the Hydraulics Division, American Society of Civil Engineers* 96: 2065–2078.

Porterfield, G. 1972. *Techniques of Water-Resources Investigations of the US Geological Survey, Book 3, Applications of Hydraulics, Chapter C3,* *Computation of Fluvial Sediment Discharge*, TWRI 3-C3, Reston, VA: US Geological Survey, 72pp.

Posada, G.L. and Nordin, C.F. 1993. Total sediment loads of tropical rivers. In: *Hydraulic Engineering '93*, vol. 1, New York: American Society of Civil Engineers, pp. 258–262.

Randle, T.J., Hilldale, R.C., Marr, J.D.G. and Johnson, S.R. 2009. *Continuous surrogate bedload measurement on the Elwha River following dam removal*. In: *Proceedings of 33rd International Association of Hydraulic Engineering and Research Congress, Vancouver, Canada*, https://www.usbr.gov/research/projects/download_product.cfm?id=192 (accessed 23 October 2015).

Reid, I., Layman, J.T. and Frostick, L.E. 1980. The continuous measurement of bedload discharge. *Journal of Hydraulics Research* 18: 243–249.

Reid, I., Laronne, J.B. and Powell M.D. 1998. Flash-flood bedload dynamics of desert gravel-bed streams. *Hydrological Processes* 12: 543–557.

Renard, K.G. and Lane, L.J. 1975. *Sediment Yield as Related to a Stochastic Model of Ephemeral Runoff*, Report ARS-S-40, Washington, DC: US Department of Agriculture, Agriculture Research Service.

Ritchie, J.C. and Cooper, C.M. 1988. Comparison of measured suspended sediment concentrations with suspended sediment concentrations estimated from Landsat MSS data. *International Journal of Remote Sensing* 9: 379–387.

Robinson, A.R. 1962. Vortex tube sand trap. *Transactions of the American Society of Civil Engineers* 127: 391–425.

Rumsby, B.A., Brasington, J., Langham, J.A., McLelland, S.J., Middleton, R. and Rollinson, G. 2008. Monitoring and modelling particle and reach-scale morphological change in gravel-bed rivers: applications and challenges. *Geomorphology* 93: 40–54.

Ryan, S.E. and Porth, L.S. 1999. A field comparison of three pressure-difference bedload samplers. *Geomorphology* 30: 307–322.

Ryan, S.E. and Troendle, C.A. 1997. Measuring bedload in coarse-grained mountain channels: procedures, problems and recommendations. In: *Proceedings of the American Water Resources Association Annual Summer Conference*, pp. 949–958.

Ryan, S.E., Porth, L.S. and Troendle, C.A. 2005. Coarse sediment transport in mountain streams in Colorado and Wyoming, USA. *Earth Surface Processes and Landforms* 30: 269–288.

Schaank, E.M.H. 1937. Discussion of: J. Smetana, Appareil pour le jaugeage du débit solide entrainé sur le fond du cours d'eau [An instrument for the measurement of bedload in rivers]. In: *Proceedings of the 1st Meeting, International Association for Hydraulic Structures Research*, Appendix 4, pp. 93–120.

Schall, J.D. and Fisher, G.A. 1996. Hydrographic surveying using global positioning techniques. In: Albertson, M.L., Molinas, A. and Hotchkiss, R., eds., *Proceedings of the International Conference on Reservoir Sedimentation*, Fort Collins, CO: Colorado State University, pp. 247–253.

Schmidt, J.C., Parnell, R.A., Grams, P.E., Hazel, J.E., Kaplinski, M.A., Stevens, L.E. and Hoffnagle, T.L. 2001. The 1996 controlled flood in the Grand Canyon: flow, sediment transport and geomorphic change. *Ecological Applications* 11: 657–671.

Seal, R., Paola, C. and Parker, G. 1993. The effect of local patchiness of gravel grain size distributions on bed load transport in braided rivers. In: Wang, S.S.Y., ed., *Advances in Hydro-science Engineering*, University of Mississippi, pp. 1331–1338.

Seydell, I. 1998. *Evaluating the Effect of Historical Land Use on Sediment Yield, Ruhama Basin, Israel*, unpublished Vertieferarbeit Thesis, Department of Civil Engineering, Technical University, Darmstadt, 57pp.

Shulits, S. and Hill, R.D. 1968. Bedload Formulas, *Agricultural Research Service Report ARS-SCW-1*, Washington, DC: US Department of Agriculture, Agricultural Research Service.

Simons, D.B., Richardson, E.V. and Nordin, C.R. 1965. *Bedload Equation for Ripples and Dunes*, US Geological Survey Professional Paper 462-H, Washington, DC: US Government Printing Office, 9pp.

Singh, A., Fienberg, K., Jerolmack, D.J., Marr, J.D.G. and Foufoula-Gergiou, E. 2009. Experimental evidence for statistical scaling and intermittency in sediment transport rates. *Journal of Geophysical Research* 114: F01025.

Singh, K.P., Durgunoglu, A. and Ramamurthy, G.S. 1986. Some solutions to underestimation of stream sediment yields. *Hydrological Science and Technology: Short Papers* 2: 1–7.

Stelczer, K. 1981. *Bed-load Transport*, Littleton, CO: Water Resources Publications, 295pp.

Sterling, S.M. and Church, M. 2002. Sediment trapping characteristics of a pit trap and the Helley–Smith sampler in a cobble gravel-bed river. *Water Resources Research* 38: 19-1–19-11.

Stevens, H.H., Jr, and Yang, C.T. 1989. *Summary and Use of Selected Fluvial Sediment Discharge Formulas*, US Geological Survey Water-Resources Investigations Report 89-4026, Denver, CO: US Geological Survey, 123pp.

Stott, T. and Mount, N. 2007. Alpine proglacial suspended sediment dynamics in warm and cool ablation seasons: implications for global warming. *Journal of Hydrology* 332: 259–270.

Sullivan, S.A. 1996. DGPS and GIS improve lake sedimentation survey procedures. In: Albertson, M.L., Molinas, A. and Hotchkiss, R., eds., *Proceedings of the International Conference on Reservoir Sedimentation*, Fort Collins, CO: Colorado State University, pp. 255–262.

Swiss Federal Authority for Water Utilization. 1939. *Untersuchungen in der Natur über Bettbildung, Geschiebe- und Schwebestoffhrung* [Field research on bed formations, bedload and suspended load movement], Mitteilung Nr. 33 des Amtes für Wasserwirtschaft, 114pp.

Tacconi, P. and Billi, P. 1987. Bed load transport measurements by the vortex-tube trap on Virginio Creek, Italy. In: Thorne, C.R., Bathurst, J.C. and Hey, R.D., eds., *Sediment Transport in Gravel-bed Rivers*, Chichester: John Wiley & Sons, pp. 583–616.

Thomas, R.B. 1985. Estimating total suspended sediment yield with probability sampling. *Water Resources Research* 21:1381–1388.

Thomas, R.B. 1988. Monitoring baseline suspended sediment in forested basins: the effects of sampling on suspended sediment rating curves. *Hydrological Sciences Journal* 33: 499–514.

Thomas, R.B. and Lewis, J. 1993a. A comparison of selection at list time and time-stratified sampling for estimating suspended sediment loads. *Water Resources Research* 29: 1247–1256.

Thomas, R.B. and Lewis J. 1993b. A new model for bed load sampler calibration to replace the probability-matching method. *Water Resources Research* 29: 583–597.

Thomas, R.B. and Lewis, J. 1995. An evaluation of flow-stratified sampling for estimating suspended sediment loads. *Journal of Hydrology* 170: 27–45.

Thompson, S.M. 1976. *Clutha Power Development, Siltation of Hydro-electric Lakes*, Wellington: Ministry of Works and Development, 39pp.

Thorne, P.D., Hardcastle, P.J. and Soulsby, R.L. 1993. Analysis of acoustic measurements of suspended sediments. *Journal of Geophysical Research* 98: 899–910.

Toffaleti, F.B. 1968. *A Procedure for Computation of the Total River Sand Discharge and Detailed Distribution, Bed to Surface*, Technical Report No. 5, Vicksburg, MS: Committee on Channel Stabilization, United States Army Corps of Engineers.

Topping, D.J., Wright, S.A., Melis, T.S., and Rubin, D.M. 2007. High-resolution measurements of suspended sediment concentration and grain size in the Colorado River in Grand Canyon using a multi-frequency acoustic system. In: *Proceedings of the Tenth International Symposium on River Sedimentation, Moscow, Russia, August 1–4, 2007*, vol. III, Moscow, Moscow State University Press, pp. 330–339.

Turowski, J.M., Badoux, A. and Rickenmann, D. 2011. Start and end of bedload transport in gravel-bed streams. *Geophysical Research Letters* 38: L044018.

Urick, R. J. 1983. *Principles of Underwater Sound*, 3rd edn., New York: McGraw-Hill, 423pp.

US Bureau of Reclamation. 2006. Reservoir survey and data analysis. In: *Erosion and Sedimentation Manual*, Denver, CO: US Department of the Interior, Chapter 9, pp. 9-1–9-65.

Vanoni, V.A., ed. 1975. *Sedimentation Engineering*, Manuals and Reports on Engineering Practice No. 54, New York: American Society of Civil Engineers, 745pp.

Vanoni, V.A., Brooks, N.H. and Kennedy, J.F. 1961. *Lecture Notes on Sediment Transportation and Channel Stability*, W.M. Keck Laboratory of Hydraulics and Water Resources Report KH-R-1, Pasadena, CA: California Institute of Technology.

Van Rijn, L.C. and Gaweesh, M.T.K. 1992. A new total sediment load sampler. *Journal of Hydraulic Engineering* 118: 1686–1691.

Van Sickle, J. 1981. Long-term distributions of annual sediment yields from small watersheds. *Water Resources Research* 17: 659–663.

Vericat, D., Church, M. and Batalla, R. J. 2006. Bed load bias: comparison of measurements obtained using two (76 and 152 mm) Helley–Smith samplers in a gravel bed river. *Water Resources Research* 42: W01402.

Wall, G.R., Nystrom, E.A. and Litten, S. 2006. *Use of an ADCP to Compute Suspended Sediment Discharge in the Tidal Hudson River, New York*, US Geological Survey Scientific Investigations Report 2006-5055, Reston, VA: US Geological Survey, 16pp.

Wallerstein, N.P., Alonso, C.V., Bennett, S.J. and Thorne, C.R. 2001. Distorted Froude-scaled flume analysis of large woody debris. *Earth Surface Processes and Landforms* 26: 1265–1284.

Walling, D.E. 1974. Suspended sediment and solute yields from a small catchment prior to urbanisation. In: Gregory, K.J. and Walling, D.E., eds., *Fluvial Processes in Instrumented Watersheds*, Institute of British Geographers Special Publication 6, London: Institute of British Geographers, pp. 169–192.

Walling, D.E. 1977. *Limitations of the Rating Curve Technique for Estimating Suspended Sediment Loads, with Particular Reference to British Rivers*, IAHS Publication No. 122, Wallingford: International Association of Hydrological Sciences, pp. 34–48.

Walling, D.E. 1984. Dissolved loads and their measurement. In: Hadley, R.F. and Walling, D.E., eds., *Erosion and Sediment Yield: Some Methods of Measurement and Modelling*, Norwich: Geo Books, pp. 111–177.

Walling, D.E. and Moorehead, P.W. 1989. The particle size characteristics of fluvial suspended sediment: an overview. *Hydrobiologia* 176/177: 125–149.

Walling, D.E. and Webb, B.W. 1988. *The Reliability of Rating Curve Estimates of Suspended Sediment Yield: Some Further Comments*, IAHS Publication No. 174, Wallingford: International Association of Hydrological Sciences, pp. 337–350.

Webby, M.G., Walsh, J.M. and Goring, D.G. 1996. Sediment redistribution in Lake Roxburgh, New Zealand. In: Albertson, M.L., Molinas, A. and Hotchkiss, R., eds., *Proceedings of the International Conference on Reservoir Sedimentation*, Fort Collins, CO: Colorado State University, pp. 1867–1881.

Wheaton, J.M., Brasington, J., Darby, S.E. and Sear, D.A. 2010. Accounting for uncertainty in DEMs from repeat topographic surveys: improved sediment budgets. *Earth Surface Processes and Landforms* 35: 136–156.

White, W.R., Milli, H. and Crabbe, A.D. 1973. Sediment Transport: an Appraisal of Available Methods*, Report 119*, Wallingford: Hydraulics Research Station.

Wilcock, P.R. 1997a. *A Method for Predicting Sediment Transport in Gravel-bed Rivers. Technical Report to the US Forest Service, Rocky Mountain Forest and Range Experiment Station*, Baltimore, MD: Department of Geography and Environmental Engineering, Johns Hopkins University, 59pp.

Wilcock, P.R. 1997b. Entrainment, displacement and transport of tracer gravels. *Earth Surface Processes and Landforms* 22: 1125–1138.

Wilcock, P.R. 1998. Two-fraction model of initial sediment motion in gravel-bed rivers. *Science* 280: 410–412.

Wilcock, P.R. 2001. Toward a practical method of estimating sediment transport rates in gravel-bed rivers. *Earth Surface Processes and Landforms* 26: 1395–1408.

Wilcock, P.R. and Crowe, J.C. 2003. Surface-based transport model for mixed-size sediment. *Journal of Hydraulic Engineering* 129: 120–127.

Wilcock, P.R. and McArdell, B.W. 1997. Partial transport of a sand/gravel sediment. *Water Resources Research* 33: 235–245.

Wilcock, P.R., Barta, A.F. and Shea, C.C.C. 1994. Estimating local boundary shear stress in large gravel-bed rivers. In: *Hydraulic Engineering '94*, New York: American Society of Civil Engineers, pp. 834–838.

Wilcock, P.R., Kondolf, G.M., Matthews, W.V.G. and Barta, A.F. 1996. Specification of sediment maintenance flows for a large gravel-bed river. *Water Resources Research* 32: 2911–2921.

Wilcock, P., Pitlick, J. and Cui, Y. 2009. *Sediment Transport Primer: Estimating Bed-material Transport in Gravel-bed Rivers*, General Technical Report RMRS-GTR-226, Fort Collins, CO: US Department of Agriculture, Forest Service, Rocky Mountain Research Station, 78pp.

Williams, D.T. and Julien, P.Y. 1989. Applicability index for sand transport equation. *Journal of Hydraulic Engineering* 115: 1578–1581.

Willis, J.C. and Kennedy, J.F. 1977. *Sediment Discharge of Alluvial Streams Calculated from Bed-form Statistics*, Iowa Institute of Hydraulic Research Report 202, Iowa City, IA: Iowa Institute of Hydraulic Research, 132pp.

Witte, W.G., Whitlock, C.H., Usry, J.W., Morris, W.D. and Gurganus, E. 1981. Laboratory Measurements of Physical, Chemical, and Optical Characteristics of Lake Chicot Sediment Waters*, NASA Technical Paper 1941*, Hampton, VA: Langley Research Center, National Aeronautical and Space Administration, 27pp.

Woidt, J.L., Cox, A.L. and Thornton, C.I. 2011. Evaluation of sedimentation and erosion trends in the Sacramento River near the M&T/Llano Seco Pump Station. In: *Proceedings of AGU Hydrology Days 2011*, Colorado State University, http://hydrologydays.colostate.edu/Papers_2011/Woidt_paper.pdf (accessed 20 October 2015).

Wren, D.G., Barkdoll, B.D., Kuhnle, R.A. and Derrow, R.W. 2000. Field techniques for suspended-sediment measurement. *Journal of Hydraulic Engineering* 126: 97–104.

Yeloff, D.E., Labadz, J.C., Hunt, C.O., Higgitt, D.L. and Foster, I.D.L. 2005. Blanket peat erosion and sediment yield in an upland reservoir catchment in the southern Pennines, UK. *Earth Surface Processes and Landforms* 30: 717–733.

Young, W.J. and Davies, T.R.H. 1990. Prediction of bedload transport rates in braided rivers: a hydraulic model study. *Journal of Hydrology (New Zealand)* 29: 75–92.

CHAPTER 16

Sediment budgets as an organizing framework in fluvial geomorphology

Leslie M. Reid[1] and Thomas Dunne[2]
[1]USDA Forest Service Pacific Southwest Research Station, Arcata, CA, USA
[2]University of California, Santa Barbara, CA, USA

16.1 Introduction

Fluvial geomorphology concerns the transport of weathered rock debris and its accumulation and organization into landforms that evolve continuously. Whereas theoretical fluvial geomorphology focuses on the mechanics of fluvial processes and the principles governing the evolution of landforms, applied fluvial geomorphology is devoted to understanding and designing strategies for coexisting with changing fluvial systems. Whenever a fluvial feature changes form, there is a local imbalance in the movement of sediment to and from the site. An understanding of how a river system collects, transports and deposits sediment is therefore central to addressing both applied and theoretical questions regarding how changes in catchment conditions affect channels, how long the effects will last and what the sequence of responses will be. Sediment budgets are tools for building that understanding.

Sediment budgets define the most fundamental aspect of landform evolution: mass conservation as it is achieved by morphogenetic processes acting within the boundary conditions imposed by natural or anthropogenic controls. Information about whether the sediment budget of a particular fluvial landform or a fluvial system is in a steady state — or the degree of imbalance if it is not — is as basic a descriptor as the mean annual flood, mean annual runoff or other commonly used characterizations of fluvial systems. Quantifying sediment supplies and transport rates is typically more difficult to accomplish than measurement of hydroclimatic quantities, but technical advances during the past few decades have improved the capacity for defining the sediment budgets of fluvial systems. Even rudimentary sediment budgets can prevent oversights and guide the selection of the analytical tools needed for more detailed analyses. Eventually, the degree of detail and sophistication of sediment budgeting is constrained by time and by the availability of data sources and tools, many of which are described in detail in other chapters of this book. This chapter discusses the nature of sediment budgets, provides examples of how they have been used and describes an approach for designing and constructing useful budgets.

The sediment budget defined

A sediment budget describes the input, transport, storage and export of sediment in a geomorphic system. For example, Fig. 16.1 encapsulates the operation of the sediment budget of a small forested mountain catchment. This budget was based initially on qualitative field observations and mapping of sediment sources and storage elements and the conceptual diagram was then used to guide quantitative estimates of the various transfer rates and storage times of the sediment.

Sediment budgets can be designed to quantify the magnitude of a process or response rate, its location and its timing or to explore the influences contributing to a morphological change. They can be used to compare the likely outcomes of different land-management options or climatic changes or to evaluate the significance and implications of climatic, tectonic or land-use changes that have already occurred. Sediment budgets provide a framework for organizing both qualitative information about process interactions and quantitative information about process rates. Budgets can take many forms, describe many scales and incorporate diverse levels of precision. The most commonly used sediment budgets take the form of qualitative flowcharts that describe relationships between sediment sources and transport processes. Long-term monitoring projects often are used to provide more precise measurements of particular budget components.

Whether qualitative or quantitative, all sediment budgets are conceptually underlain by the continuity equation for sediment transfer:

$$\text{sediment input to a landscape element} = \text{sediment output} + \text{change in sediment storage} \quad (16.1)$$

where all terms are expressed as quantities per unit time. The basic equation can be refined in many ways. Changes in grain

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

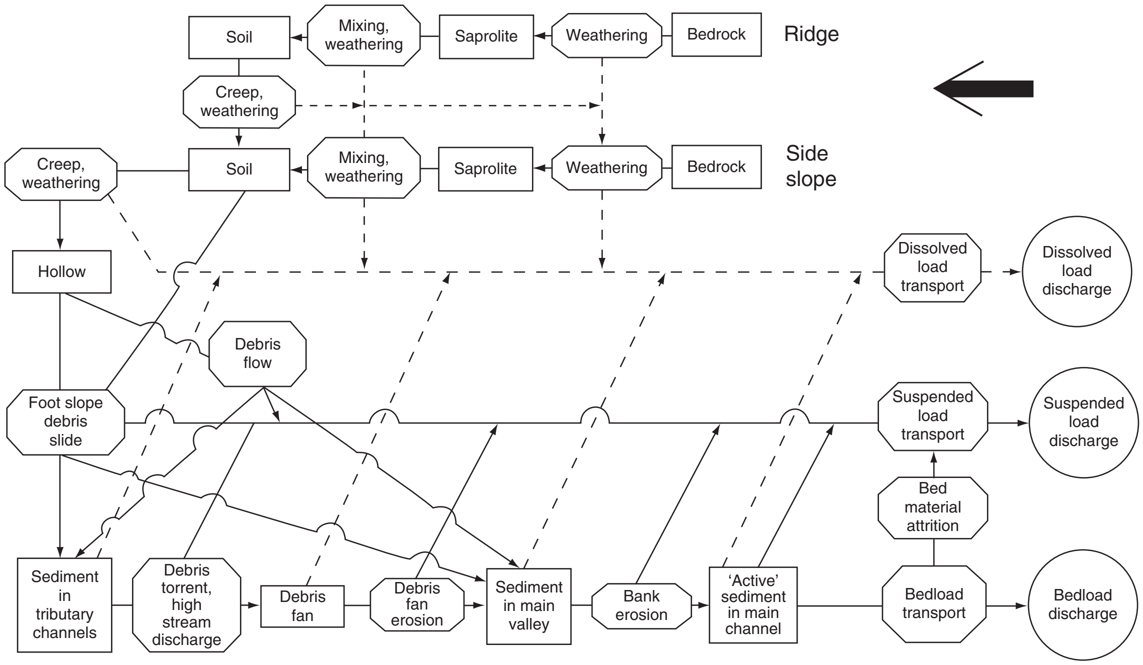


Figure 16.1 Conceptual model of the sediment budget of a small mountainous watershed in the Oregon Coast Range. Rectangles represent storage elements, octagons indicate transfer processes and circles represent outputs. Solid lines indicate the transfer of sediment and dotted lines represent the migration of solutes. Dietrich and Dunne, 1978. Reproduced with permission of Schweizerbart.

size can be accounted for by constructing the equation for different size classes, for example, and specific processes can be isolated.

Given the various forms that sediment budgets may take and the variety of problems to which they can be applied (Table 16.1), it is clearly not useful to think of sediment budgeting as a single tool to be applied using a uniform protocol. Instead, sediment budgeting represents a general approach to geomorphic problem solving and the methods most useful for each budget depend on the intended application of that budget.

History and applications

Geomorphologists have long used the concept that imbalances between sediment supply and transport capacity cause aggradation and degradation, and results of sediment production and transport measurements were being used to understand landscapes by the late 1800s. Hill (1896), for example, integrated the results of landslide surveys in New Zealand to demonstrate that landslides could influence landscape evolution. Gilbert (1917) used a sediment budget to evaluate the impacts of hydraulic mining in California on downstream navigation, and although later work in the same river system added detail (James 1997; Singer *et al.* 2013), Gilbert's basic insight remains useful.

Sediment budgeting as a concept for interpreting sparse data soon proved useful for analysing the landscape-scale effects of land use. Both Haggett (1961) in southeastern Brazil and Trimble (1977) in the Southern Appalachian Mountains found large disparities between landscape-averaged estimates of soil erosion following European colonization and subsequent amounts of fluvial sediment transport in neighbouring lowlands. Both authors interpreted the disparities to indicate that large volumes of sediment must be stored on footslopes and valley floors and would continue to contribute fluvial sediment long after hillslopes restabilize. These studies emphasized the connections between sediment fluxes through landscape elements and highlighted the importance of changes in sediment storage. Although these sediment budgets were rudimentary by modern standards, they revealed the significance of long-term, intermittent transfer and storage of sediment throughout landscapes and stimulated decades of research.

By the mid-1900s, methods had been developed to quantify components of sediment regimes and long-term monitoring records and aerial photographs were becoming available. The idea of systematically quantifying the balance between sediment inputs, transport rates and storage changes began to spread. At first, sediment budgets simply involved systematic accounting of process measurements (for a review, see Reid and Dunne 1996). More recently, field monitoring studies were coupled with modelling to interpret, extend and generalize results. New methods of dating now allow long-term deposition rates to be evaluated for large sediment sinks and analyses of cosmogenic isotope concentrations in sediment are used to infer average erosion rates over large catchments and regions.

As methodological and conceptual difficulties were surmounted, the organizing power of the sediment budget concept became more evident and the approach is now widely applied to quantify landform evolution under both natural and modified

Table 16.1 Examples of sediment budgets used to address issues in fluvial geomorphology.

Problem and reference	**Why** PDF	**Regime** DESY	**Precision** LNS	**Time** AELS	**Method** MFAESH
Spatial focus: Catchment response					
Prioritize rehabilitation by erosion potential (Gellis *et al.* 2001)	.D.	.E..	..S	...S	.FA...
Proportion of sediment yield from landslides (Hovius *et al.* 1997)	.D.	.E..	.N.	A...	.FA...
Sediment contribution to lake from cyclone (Page *et al.* 1994)	.D.	DESY	.N.	.E..	.FA.S.
Effectiveness of soil conservation strategies (Phillips 1986)	..F	DESY	.N.	..L.	.F.
Downstream influence of upper-basin sediment (Phillips 1991)	PD.	DESY	.N.	..L.	.F.
Relation of hillslope erosion to sediment yield (Reneau and Dietrich 1991)	P..	.E.Y	.N.	A.	F..SH
Spatial focus: Channel system response					
Plan restoration using erosion distribution (Abernethy and Rutherfurd 1998)	.DF	DE..	L.S	S	.F.E..
Extent of channel recovery from old mining debris (James 1997)	.DF	DES.	..S	..L.	MFA..H
Effect of mining on downstream channels (Knighton 1991)	P.F	DESY	.N.	..L.	.FA
Long-channel trends in sediment character (Le Pera and Sorriso-Valvo 2000)	.D.	D	.N.	S	.F.
Effect of land use on downstream channel form (Liébault and Piégay 2001)	.D.	.ES.	.N.	..L.	.FAES.
Extent of channel recovery from a major flood (Madej and Ozaki 1996)	P..	DESY	.N.	.E..	MFA
Downstream distribution of mining debris (Marron 1992)	P..	DES.	.N.	..L.	.FA.S.
Effects of land use on downstream conditions (Trimble 1983)	P..	DESY	.N.	..L.	.F.E..
Develop strategy for catchment rehabilitation (Trimble 1993)	..F	DESY	..S	..L.	.F.E..
Spatial focus: Response of a particular reach					
Cause of change in channel form (Brooks and Brierley 1997)	P..	.ES.	L..	..L.	.F..S.
Effect of gravel mining on channel form (Collins and Dunne 1989)	P..	.ES.	.N.	..L.	.FAE.H
Design appropriate gravel harvest rate (Davis *et al.* 2000)	.D.	DES	..S	S	.F.E.H
Sediment exchanges between channel and floodplain (Dunne *et al.* 1998)	.D.	DESY	.N.	A	MFAE..
Describe original river sediment regime (Kesel *et al.* 1992)	P..	.ESY	.N.	A	..H
Particle transport mode variation through a reach (McLean *et al.* 1999)	.D.	D.SY	.N.	A	M..
Effect of channelization on a wetland (Nakamura *et al.* 1997)	P..	D.S.	.N.	..L.	MFA
Effect of dam on downstream sediment load (Phillips *et al.* 2004)	.D.	DESY	.N.	..L.	.FAE.H
Manage river to improve fish habitat (Pitlick and Van Steeter 1998)	..F	DES.	.N.	..L.	MF.E..
Extent of sand deposition during floods (Ten Brinke *et al.* 1998)	.D.	D.S.	.N.	.E..	.FA
Extent of deposition of suspended sediment load (Walling *et al.* 1998)	.D.	..S.	.N.	S	MF..S.
Design dam release regime for bed material (Wilcock *et al.* 1996)	..F	.ES.	.N.	A	E..
Downstream effect of sediment release (Wohl and Cenderelli 2000)	.D.	DESY	.N.	.E..	MF.
Effect of dams on Yangtze delta (Yang *et al.* 2005)	.DF	.ESY	.NS	S	..A..
Spatial focus: Specific land use, landform, etc.					
Controls on gully form and evolution (Harvey 1992)	PDF	DESY	.N.	A	MFA
Sediment input during road construction (Megahan *et al.* 1986)	.D.	.E.Y	.N.	..L.	M..
Effect of land use on lake sedimentation (Page and Trustrum 1997)	P..	Y	.N.	A.L.	.S.
Effect of small dams on national sediment budget (Renwick *et al.* 2005)	.D.	..SY	.N.	A	E.H

Why: Purpose
P Explains past development
D Describes present system
F Forecasts future conditions

Component of sediment regime
D Involves spatial distribution
E Evaluates erosion
S Evaluates sediment storage
Y Evaluates sediment yield

Precision
L Qualitative
N Quantitative
S Semiquantitative (e.g. rankings)

Time considered
A Generalized or long-term average
E Effect of a specific event
L Selected to evaluate land use
S Referenced to specific period

Method
M Monitoring carried out for the study
F Field measurements or observations
A Aerial photograph interpretation
E Modelling or published equations
S Analysis of sediment deposits
H Historical records or archived data

conditions (Table 16.1). Sediment budgets now play a key role in basic and applied geomorphological studies over a wide range of scales and levels of complexity. For example, Flemings and Jordan (1989) used a model of mountain building, isostasy and crustal flexure to analyse the partitioning of sediment between an evolving orogen, the adjacent sedimentary basin and export downstream, and Church and Slaymaker (1989) illustrated the importance of lagged and indirect responses in erosion and sedimentation during and after glaciation. Questions about the response of rivers to perturbations such as land use (Trimble 1974), dam construction and gravel mining (Kondolf and Swanson 1993) and sea-level rise (Allison *et al.* 1998) have also been explored by systematically accounting for input and output of sediment. Other studies have examined the exchange of sediment between channels and their floodplains (Marron 1992; Dunne *et al.* 1998). More recently, budgets have been used to predict effects of climate change (Lane *et al.* 2007) and to investigate the processes of carbon cycling (Cole *et al.* 2007).

Because sediment affects many ecosystem and watershed processes, sediment budgets can also be used to explore biogeochemical issues. Graf (1994) and Malmon *et al.* (2002), for example, studied the migration of radionuclides through channels and floodplains of Los Alamos Canyon, New Mexico, noting in particular the disparate trajectories of coarse sediment that contains little contaminant and the more reactive fine sediment. Walling *et al.* (2003) evaluated the role of floodplain sedimentation on contaminant flux along several rivers in northern England and Singer *et al.* (2013) quantified the role of sediment exchanges in distributing mercury within the Central Valley of California. The fate of carbon in large river systems is another emerging target of sediment budgeting (Aufdenkampe *et al.* 2011)

Sediment budgeting has contributed to the management of sediment-related problems. Studies have described the effects of logging on sediment regimes through long-term monitoring (Swanson *et al.* 1982) and have quantified the effects of specific activities such as road construction (Megahan *et al.* 1986) and road use (Reid and Dunne 1984), providing information useful for targeting sediment control efforts. Sediment budgeting has been used to design strategies for catchment-scale sediment control (Phillips 1986; Trimble 1993; Gellis *et al.* 2001) and riparian restoration (Abernethy and Rutherfurd 1998) and to plan reservoir releases to maintain habitat for particular species (Wilcock *et al.* 1996; Pitlick and Van Steeter 1998). Sediment budgets for river channels have guided the establishment of appropriate gravel extraction rates (Collins and Dunne 1989; Davis *et al.* 2000).

Dams exert a growing influence on sediment transport regimes and downstream channel responses that impact valued resources or infrastructure. Sediment budgeting provides a tool for understanding the causes and long-term outcomes of such impacts by accounting for magnitudes and spatial distributions of sediment supplies and also changes in sediment transport, deposition and erosion (Phillips *et al.* 2004; Yang *et al.* 2005; Vericat and Batalla 2006). Impoundments remove sediment from rivers and the resulting unsatisfied transport capacity of the emerging flow often causes extensive bed degradation downstream (Williams and Wolman 1984). The extent of the degradation, however, depends on how rapidly the catchment downstream augments the supply of bed material load. In some cases, the reduction of transport capacity due to decreases in flood peaks and the supply of sediment from the undammed catchment allow the channel bed to aggrade within a short distance of the dam, beginning with fans or bar accumulations at tributary mouths.

Sediment budgets are increasingly used to aid regulatory oversight of land-use activities. Budgets have been used to develop 'total maximum daily load' allocations and sediment control plans required by the US Clean Water Act for non-point-source sediment in impaired catchments. Sediment budgeting can also aid the assessment of environmental impacts from planned projects. Downstream cumulative impacts, in particular, often result from changes in erosion, transport or deposition of sediment.

16.2 Understanding and assessing components of the sediment system

A sediment system can be examined from many points of view and each of these could be represented by a sediment budget. Which point of view is most useful depends on the intended application. To understand the variety of approaches possible and the analytical challenges they involve, the components of a catchment's sediment production and transport system must first be understood. A wealth of literature is available about specific aspects of the sediment system and other chapters in this book discuss sediment transport and channel change. Here we summarize concepts that are particularly relevant to sediment budgeting and describe assessment methods applicable to components of the sediment system. Samples of relevant references describing applications of the concepts are included in Tables 16.2 and 16.3.

Hillslope processes and sediment delivery to streams

Sediment in a catchment originates from bedrock, atmospheric deposition and biological activity. Bedrock becomes sediment through physical and chemical weathering, during which some of the original material is removed by dissolution. The 'soil production rate' (Heimsath *et al.* 1997) or 'regolith production rate' (Small *et al.* 1999) is the rate per unit area at which soil material is converted from bedrock.

As weathering progresses, a particle may remain in place as saprolite or be dislodged ('eroded') and transported downslope as colluvium. Erosion rates are generally described as a net loss of sediment per unit area or a rate of surface lowering, whereas transport rates represent the discharge of sediment per unit width of hillslope or through a channel cross-section. Net erosion occurs only where transport into an area is less than transport out.

Ordinarily, a sequence of disparate hillslope processes (Table 16.2) moves sediment particles intermittently downslope to a channel. The rate of sediment production to stream channels has been defined as the rate of colluvial sediment transport across a line corresponding to the stream bank (Reid and Dunne 1996). The words 'production' and 'delivery' can refer to transfer between any landscape elements, so the context for the usage must be considered carefully to avoid confusion or double-accounting.

In an accounting of primary sediment input, any particle can be delivered to the stream system only once. For example, soil creep moves sediment to the base of a slope, where bank erosion carves away the encroaching sediment. In this case, sediment

Table 16.2 Examples of methods used to evaluate erosion, colluvial sediment transport and primary sediment production to channels. Major controlling variables in parentheses. Expected accuracies estimated for typical conditions: **H**, 0.6–1.6 times actual; **M**, 0.4–2.5 times actual; **L**, <0.4 or >2.5 times actual, increasing with more detailed work or long-term monitoring and decreasing if reconnaissance methods used. References provide further information or examples, which can be expanded with bibliographic research.

Examples of analysis methods	References
Dissolution (topography, climate, bedrock, soil depth, vegetation)	
Monitor volume and concentration of lithogenic solutes in streamflow and precipitation. Relation between concentrations and specific conductivity, which is readily monitored. Concentration-streamflow may vary seasonally and by solute; apply to the annual hydrograph to calculate annual lithogenic solute yield; subtract inputs from precipitation (**H**)	Janda (1971); Dunne (1978); Anderson and Dietrich (2001)
Soil creep (gradient, climate, soil type, soil depth, vegetation)	
Difficult to monitor; few measurements exist. Method 1: apply values of creep rate measured at similar sites and multiply estimated creep discharge per unit width by the length of colluvial stream bank (**L**). Method 2: estimate creep transport from processes such as bank erosion and stream bank landslides, which are supplied by it (**M**)	Saunders and Young (1983); Auzet and Ambroise (1996); Reid and Dunne (1996)
Burrowing (gradient, species, soil type, vegetation)	
Production is by transport of excavated sediment across stream banks; measure deposit volumes, considering seasonal distributions (**H**). Delivery by overland flow possibly important. Must know burrow patterns to assess on-slope transport (**L**)	Hall *et al.* (1999); Gabet *et al.* (2003)
Tree-throw (gradient, storm size, vegetation type and age)	
Identify uprooting density for each vegetation type and use age of associated vegetation to identify fall-age diagnostics (e.g. time to shedding of twigs or loss of bark). Field sample to estimate delivery ratios and number of contributing rootwads by age per unit channel length (**H**). For transport rate, sample frequency per unit area, rootwad volumes (minus root volume) and displacement of mounds from scars; consider wind storm history (**H**)	Schaetzl *et al.* (1989); Norman *et al.* (1995); Gallaway *et al.* (2009)
Earthflows (gradient, seasonal rainfall, bedrock, vegetation)	
Map flows on aerial photographs; less visible flows require fieldwork. Delivery is by bank erosion, gully erosion and shallow landsliding. Method 1: use methods described below to assess rates of delivery processes (**M**). Method 2: estimate surface velocity near toe from displacement of survey markers or features visible on sequential air photos, assume a characteristic velocity profile (or measure using inclinometer tubes) and apply the resulting unit discharge to the measured flow cross-section (**H**)	Van Asch and Van Genuchten (1990); Zhang *et al.* (1991); Nolan and Janda (1995); Roering *et al.* (2009)
Deep-seated landslides (gradient, seasonal rainfall, bedrock, vegetation)	
Map and date using sequential aerial photographs; field sample to measure sediment delivery (compare scar and deposit volumes), evaluate slides not visible on photographs and date those scars using vegetation. Assess as for earthflows if movement is chronic or intermittent; if removal of deposits is intermittent, evaluate temporary storage. Calculate production as frequency × volume × delivery ratio for each land stratum. Consider recent rainfall patterns when interpreting average rates (**H**)	Ibsen and Brunsden (1996); Corominas and Moya (1999); Korup (2005); Schuerch *et al.* (2006); Schwab *et al.* (2008)
Shallow landslides (gradient, landform, storm rainfall, bedrock, vegetation, earthquakes)	
Map and date using sequential aerial photographs; field sample to measure sediment delivery (compare scar and deposit volumes), identify slides not visible on photographs and date them using vegetation. May be able to define relationships between scar area and volume and between topographic setting and delivery ratio. Calculate production as frequency × volume × delivery ratio for each land stratum. Consider rainstorm history when interpreting average rates (**H**)	Mantovani *et al.* (1996); Hovius *et al.* (1997); Reid (1998); Corominas and Moya (1999); Gabet and Dunne (2002); Malamud *et al.* (2004)
Debris flow erosion (hillslope gradient, storm rainfall, channel gradient, bedrock, vegetation)	
Under steady state, only erosion of colluvium and bedrock is primary; otherwise, also evaluate remobilization of ‘legacy’ channel deposits. Map and date visible scars using aerial photographs; measure widths in the field if obscured by trees. Identify other flows in the field from debris deposits and date using vegetation. Determine characteristic erosion depths from scarp heights and from depths of soil and channel deposits at analogous sites. Consider storm history when interpreting average rates (**H**)	Van Steijn (1996); Cenderelli and Kite (1998); Santi *et al.* (2008); Stoffel *et al.* (2008); Guthrie *et al.* (2010)
Primary streambank erosion (gradient, peak-flow size, channel size, soil type, vegetation)	
Under steady state, only colluvial and bedrock erosion is primary; otherwise, evaluate erosion of ‘legacy’ alluvium to assess storage changes. Stratify by channel type. Estimate bank retreat rates in large channels using sequential air photos and field measurements of bank height (**H**). Otherwise, field sample to estimate proportion of banks eroding. Rates often difficult to assess without monitoring, but might be estimated from datable vegetation, scarp depths or deposit volumes (**M**)	Kesel *et al.* (1992); Barker *et al.* (1997); Stott (1997); Couper and Maddock (2001); Pizzuto *et al.* (2010)
Primary channel erosion (channel gradient, peak flow size, channel size, bedrock)	
Under steady state, only colluvial and bedrock erosion is primary. Channel area is small relative to hillslope area so long-term, steady-state production from this source is relatively small. If steady state cannot be assumed, evaluate incision of alluvium also (Table 16.3); terrace surfaces often can be dated to calculate incision rates after surfaces formed. Channel cross-section monitoring (**H**)	Trimble (1997); Ward and Carter (1999); Reneau (2000); Bishop *et al.* (2005); Stock *et al.* (2005); Cook *et al.* (2009)
Tunnel erosion (gradient, landform, soil type)	
Examine channel heads to ascertain presence of tunnels (‘soil pipes’); identify upslope extent by probing, trenching or observing collapse scars. Sediment delivery is most reliably assessed by monitoring effluent from multiple tunnels (**M**)	García-Ruiz *et al.* (1997); Sayer *et al.* (2006)

(*continued overleaf*)

Table 16.2 (*continued*)

Examples of analysis methods	References
Gullying (gradient, catchment area, peak flow size, channel size, soil type, vegetation, compaction)	
Map and date gullies in open terrain using sequential aerial photographs; construct relations between length or area and volume from field measurements and use these to estimate volume changes through time (**H**). Otherwise, field sample for distribution, frequency, size and age. Date using associated vegetation, eye-witness accounts or age of causal features. Estimate sediment production through time from headward retreat rate and volume–length relationships (**H**)	Nachtergaele and Poesen (1999); Vandekerckhove *et al.* (2003); Ghimire *et al.* (2006); Nyssen *et al.* (2006); Giménez *et al.* (2009)
Rilling (gradient, slope length, storm rainfall, soil type, vegetation, compaction)	
Assess distribution considering controlling variables and season. Monitor or field sample rill dimensions before and after wet season or storms of different sizes or through year. Estimate delivery ratio from size distribution and volume of deposits (compare with soil texture) or from sediment concentration measurements (**H**). Widely used surface erosion equations usually include both rill and sheetwash erosion	Collins and Dunne (1986); Moody and Martin (2001)
Sheetwash erosion (gradient, slope length, storm rainfall, soil type, vegetation, compaction)	
Define distribution by controlling variables and season. Field sample root exposure on datable plants or monitor using erosion pins (**H**). Such measurements combine effects of sheet, rainsplash, dry ravel and wind erosion, so use the spatial and temporal distribution of each to interpret results. Estimate delivery ratio from size distribution and volume of deposits (compare with soil texture) or from sediment concentration measurements. If using an erosion equation, test by comparing predictions against monitoring data (even short-term data can indicate whether results are reasonable) or other field evidence (**H**). Surface erosion can also be quantified by sampling runoff from small, definable catchments (**H**)	Reid and Dunne (1984); Collins and Dunne (1986); Yanda (2000); Merritt *et al.* (2003); Bodoque *et al.* (2005); Kinnell (2010)
Dry ravel (gradient, soil moisture, soil type, temperature, vegetation)	
Define distribution by controlling variables and season, including relationship to fires. Measure root exposure on datable plants or monitor erosion pins or accumulation in troughs. Such measurements combine effects of sheet, rainsplash, dry ravel and wind erosion, so consider their spatial and temporal distributions to interpret results. Compare grain sizes of deposits and sources to estimate delivery (**H**)	Megahan *et al.* (1983); Gabet (2003); Jackson and Roering (2009); DiBiase and Lamb (2013)
Construction, tillage, engineering, etc. (gradient, type of project, soil type, bedrock)	
Only direct mechanical displacement of sediment is considered here; secondary processes are considered above. Aerial photographs, maps, plans, interviews with equipment operators and field observations can indicate distribution and timing of effects and location of displaced material with respect to streams (**H**). Monitor sediment washed from definable mini-catchments or plots **(H)**	Reid and Dunne (1984); Phillips *et al.* (1999); Zhang *et al.* (2004); Marden *et al.* (2006); Van Oost *et al.* (2006)
Deposition on hillslopes and swales (gradient, landform, soil type, vegetation, sediment input)	
For non-discrete processes, field sample accumulation depths around datable plants or structures or measure seasonal accumulations atop leaf litter (**H**). Stratigraphic dating methods can be used for long-term accumulations (**H**). For discrete processes, measure volumes of deposits and date using sequential aerial photographs or ages of associated plants (**H**)	Page *et al.* (1994); Vandaele *et al.* (1996); Beuselinck *et al.* (2000); Nearing *et al.* (2005); Descheemaeker *et al.* (2006); Notebaert *et al.* (2009)

delivery is evaluated from either the rate of stream bank erosion or the soil creep discharge at the channel margin, but these rates cannot be summed because both processes involve the same particles. Similarly, sediment production cannot be calculated by summing rates of landsliding and soil creep where creep transports colluvium to bedrock hollows that are episodically evacuated by landslides (Reneau and Dietrich 1991; Dunne 1998). A portion of the sediment derived from colluvium may be deposited downstream and later re-enter the channel through erosion of alluvium. Such re-entry (described in Table 16.3) represents remobilization from temporary storage rather than primary sediment delivery. In contrast, sediment introduced by channel incision into bedrock or colluvium represents primary sediment delivery (Table 16.2).

Colluvial transport processes are of two kinds. 'Chronic' processes include transport by rainsplash, soil creep, sheetwash and other mechanisms that recur frequently at the same sites. 'Discrete' processes, in contrast, are localized events that can be counted, such as landslides and tree-throws. Table 16.2 describes methods for measuring a selection of process rates and attributes and provides examples of studies that have evaluated each process. Other methods, such as long-term monitoring, are also available for most processes listed.

Chronic processes are usually evaluated by determining average rates and applying those to the areas affected. For example, the rate of sheetwash erosion on rangelands might be estimated from measurements of root exposure around datable vegetation, by monitoring surface lowering at stakes or with web-accessible models such the USLE, WEPP, KINEROS2 or EUROSEM, suitably validated or calibrated against some empirical information. The estimated rate would then be assumed to represent the area of similar land-use activity, topography and soil type. Only part of the eroded sediment is delivered to channels, however, and this amount varies with the conditions in and around the eroding sites. The delivery ratio can be estimated for different site types using methods such as monitoring sediment transport in overland flow during a few storms, applying erosion models or comparing the grain size distribution of deposits

Table 16.3 Examples of methods used to evaluate sediment transport and storage in channels, erosion of alluvial sediment and sediment yield. Major controlling variables listed in parentheses. Expected accuracies estimated for typical conditions: **H,** 0.6–1.6 times actual; **M**, 0.4–2.5 times actual; **L**, <0.4 or >2.5 times actual. Accuracy increases with more detailed work or long-term monitoring and decreases if reconnaissance methods are used. References selected to provide further information or examples.

Examples of analysis methods	References
Bedload (channel gradient and form, flow distribution, grain size, sediment input, bedrock)	
Where coarse load is trapped in a lake or low-gradient reach, estimate transport by measuring temporal changes in depositional landforms using topographic surveys or aerial photographs (**H**). Bedload sampling data are available for a few stations, but records are usually sparse and short. Otherwise, use carefully selected bedload transport equations, appropriate for the conditions being assessed (**M**)	Reid and Dunne (1996); McLean and Church (1999); Davis *et al.* (2000); Brasington *et al.* (2003); Pelpola and Hickin (2004); Wilcock *et al.* (2009)
Suspended load (channel gradient and form, flow distribution, grain size, sediment input, bedrock)	
Measure suspended sediment concentrations over a range of flows to define a sediment rating curve and apply the resulting curve to annual hydrographs (**M** to **H**). Sediment transport equations for suspendable bed-material load are useful if input-dependent washload is not large (**M**)	Reid and Dunne (1984); Asselman (2000); Moatar *et al.* (2006); Gao (2008); Wang *et al.* (2009)
Sediment attrition (transport rate, grain size, rock type)	
Tumbling-mill experiments can indicate grain-size changes per unit travel distance (**H**). If different lithologies are present in bed material, use changes in relative abundance to estimate relative breakdown rates (**M**)	Kuenen (1956); Collins and Dunne (1989); Lewin and Brewer (2002); Le Pera and Sorriso-Valvo (2000)
Bed aggradation (channel gradient and form, flow distribution, grain size, sediment load)	
Land surveys or surveys for bridge planning can be repeated; local residents can describe recent changes; and engulfed artefacts, woody debris or plants can indicate the extent and timing of aggradation, as can changes in overbank flood severity. Long-term flow gauging data can document changes in bed elevation. Recently aggraded bed material often is finer grained and can be probed to determine the depth to a coarser gravel layer. Establish timing from personal accounts, vegetation ages and comparison of sequential aerial photographs. Estimate bar aggradation rates by multiplying the areas of bars deposited by average bar heights (**H** to **M**)	Brooks and Brierley (1997); Wathen and Hoey (1998); Lisle and Hilton (1999); Sloan *et al.* (2001); Faustini and Jones (2003); Lancaster and Casebeer (2007)
Floodplain aggradation (channel gradient and form, flow history, grain size, sediment load, vegetation)	
Measure deposit depths around datable plants or structures or date deposits using methods described in other chapters (**H**). Data from sediment traps or stakes can indicate relation between deposition and flood size, as can post-flood observations of deposition; data from large floods are needed to estimate long-term rates (**H**). Many methods for assessing bed aggradation can be applied to banks and floodplains. Several-decade-long cores can be dated with ^{137}Cs concentration profiles, profiles of ^{210}Pb attached to clay particles can provide longer records and ^{14}C dating produces records dating back thousands of years	Ten Brinke *et al.* (1998); Gomez *et al.* (1999); Rumsby (2000); Lecce and Pavlowsky (2001); Knox (2006); Aalto *et al.* (2008); Hoffmann *et al.* (2009); Provansal *et al.* (2010)
Channel erosion of alluvial sediments (channel gradient and form, flow distribution, grain size, sediment load, bedrock)	
Compare channel geometry to that of unaffected channels (**H**). Land surveys or cross-sections surveyed for bridge planning can be resurveyed if available. Calculate river-bed elevation trends at gauging stations from low-flow stage records and flow–depth measurements. Residents can describe recent changes and undercut vegetation or exposed bridge piers may provide data. Timing is usually established from personal accounts or comparison of sequential aerial photographs. Evaluate erosion rates from shifting of large channels by multiplying the areas of bank eroded by the average bank height. (**H** to **M**)	James (1997); Gonzalez (2001); Liébault and Piégay (2001); Miller *et al.* (2001); Kesel (2003)
Sediment yield (catchment size, flow distribution, sediment input, bedrock, vegetation, topography)	
Where catchments drain into lakes or ponds, yield can be estimated from rates of lake sedimentation if the trap efficiency is known and the bathymetry has been monitored or can be reconstructed (**H**). Measurements or calculations of sediment transport at the mouth of a catchment provide an estimate of yield (**H** to **M**). Nearby catchments with similar characteristics are expected to have similar sediment yields (**M**)	Wilby *et al.* (1997); Lloyd *et al.* (1998); Verstraeten and Poesen (2002); Tamene *et al.* (2006).

with that of the eroding material. Sediment production rates are then calculated by multiplying hillslope sediment yields by sediment delivery ratios for each site type and applying these values to the distribution of site types present. The parameter values of predictive models, however, are known only very approximately, despite thousands of plot-years of observations. Most applications of such models result only in discrimination of areas which produce large amounts of sediment and those which provide little. Nevertheless, such discrimination is often sufficient for highlighting which processes or landscape components dominate the sediment supply.

Rates of discrete processes, such as landslides, usually are evaluated by applying the measured spatial and temporal frequency of events to the area susceptible. Shallow landslide scars, for example, ordinarily are counted on sequential aerial photographs to determine the number of slides per unit area per unit time. Fieldwork is usually necessary to define a relationship between scar area and scar volume and to determine the proportion of landslide debris characteristically delivered to streams, and is also useful for estimating the frequency of landslides too small to detect on photographs. Because shallow landslides are generally triggered by infrequent, large storms, the dependence of areal landslide density on the magnitude of triggering events may need to be defined to determine whether the sampling period is long enough to estimate valid average rates. For many applications, only the relative rates between different land uses or landforms need be known and results from a single extensive storm often can provide this information.

Analysis of other process rates generally follows similar patterns (Table 16.2). The success of each rate analysis depends on (i) having a well-defined objective that identifies the information required, (ii) using a sampling design that permits valid characterization of the process and (iii) recognizing the area and time period over which the estimate applies. Wherever possible, rates should be estimated using multiple methods and should be checked for consistency; this is particularly important if rates are to be modelled in areas or under conditions for which the model has not been adequately tested.

Sediment transport in channels

Changes in hillslope sediment transport arouse concern when sediment reaches a channel. Incoming sediment can modify a channel's bed, morphology and sediment transport rates and may change the dominant transport mode. Streams transport sediment in three ways. The largest grains are rolled or jostled along the bed as 'bedload', while the smallest particles are continuously suspended in the flow ('washload'). Intermediate grains are entrained repeatedly by eddies and move predominantly as suspended load. These intermediate sizes return to the bed when flow slows and are referred to as 'bed material suspended load'. Most sediment in the streambed represents size fractions moved as bedload (especially in gravel-bed channels) or bed material suspended load (in sand-bed channels). The transport mode for a particular grain varies with flow and with channel characteristics.

Travel times for different components of the sediment load vary widely. Washload can exit a 1500 km^2 catchment during the same storm that eroded the sediment from a headwater hillslope, while bedload particles may require many decades to move the same distance. Typical long-term average annual travel distances for particles that are stored intermittently in channel beds are 100s of metres per year for gravel in small streams, 100–1000s of metres per year for gravel in large braided rivers, and 100–1000s of metres per year for sandy bedload (see studies described by Bunte and MacDonald 1999). Matisoff *et al.* (2002) demonstrated that concentrations of the isotopes ^{7}Be, ^{137}Cs and ^{210}Pb in suspended sediment can be used to measure the speeds and distances of fine sediment transport in single flood seasons.

The most accurate estimates of sediment transport rates in channels are provided by well-designed networks of monitoring stations with records long enough to produce representative results. However, most sediment budgets must be constructed too rapidly for such monitoring to be useful unless the data already exist and can be generalized statistically. In those cases, rates can be estimated using empirical or calibrated transport equations. Transported amounts can sometimes be obtained by measuring the volume of sediment deposited in natural or artificial sediment traps (such as alluvial fans or reservoirs) over a known period (Table 16.3).

In the absence of direct measurements, theoretical transport equations can provide useful estimates of non-washload components if the equations were calibrated over the range of conditions needed for the application. Reid and Dunne (1996) published comparisons between predicted and observed results and identified equations that appear to be reliable for various bed materials and channel sizes. Results are usually more accurate for sand-bedded than for gravel-bedded channels, but even the most reliable equations generally are accurate only to within a factor of two. Because washload is influenced more by sediment availability than by flow properties, transport equations are not useful if this component is important to the problem at hand. Instead, short-term monitoring results can be used to produce sediment rating curves, which can then be combined with calculated or measured hydrographs to estimate total suspended sediment loads, but they can be misleading if high flows are not sampled. As with any monitoring-based method, errors are introduced if the monitoring period is unrepresentative or if estimates are made by extrapolation beyond the conditions measured. However, if applied carefully, the method is the best available for predictions of washload and probably of all suspended load.

During transport, sediment grains are subject to fracture, abrasion, weathering and dissolution, contributing to widely observed downstream decreases in grain size and shifts in particle composition. Downstream fining is also influenced by size-dependent transport and additions of sediment along the channel, so attrition rates are not directly calculable from downstream size trends. Breakdown rates have also been estimated by measuring changes in clast size distribution as a function of 'travel distance' in rock tumblers that have been modified to provide realistic rates of particle interaction.

Channel and floodplain sediment storage

Periods of significant sediment transport in channels are interspersed with much longer periods when most of the sediment is temporarily stored in the channel bed, bars and floodplains. Durations of temporary storage vary by depositional feature and by location in a catchment. Small amounts of even washload-size sediment can be trapped within the bed or bank material during transport or can infiltrate as flows recede and fine sediment carried over banks can settle quickly onto floodplains. Storage on floodplains is favoured in rivers with high concentrations of particularly fine-grained sediment. Sediment deposited on floodplains generally remains in place until eroded by channel migration, so its residence time is greater where channel migration is slow.

Clay, silt and fine sand are also deposited on stream banks. Residence times can be very long if banks are well vegetated and sediment is remobilized only by bank erosion, but slumping and rilling as flows recede can reintroduce some of the newly deposited sediment. Silt and sand can accumulate on stream beds if sediment loads are particularly high or transport capacity is perennially or seasonally low. Where aggradation is triggered by an altered balance between input and transport capacity, pools commonly fill first (e.g. Lisle and Hilton 1999; Wohl and Cenderelli 2000).

Gravel is usually deposited within the channel and incorporated into the floodplain as the channel migrates. However, unless the channel is aggrading, most coarse sediment resides in bars until the next bed-mobilizing flow moves the clasts further downstream (Hassan *et al.* 1991). Large increases in coarse sediment inputs to a channel network can produce temporary waveforms of gravel that can be either mobile or stationary (Jacobson and Gran 1999; Lisle *et al.* 2001).

No landscape is unchanging, but many change slowly enough that 'steady-state' rates of sediment production and deposition can be assumed over useful time-scales. Over the long term, the evolution of landforms alters rates (e.g. erosion rates may decrease as progressive erosion reduces hillslope gradients), whereas over a shorter period, weather patterns would need consideration (e.g. 10-year-old flood deposits may provide a temporary sediment source). On average, however, if no areas of chronic aggradation or incision exist downstream, sediment contributed to a stream system under steady-state conditions roughly balances the sediment exported from the catchment.

In catchments with rapidly evolving landforms or changing conditions, this simplified view must be expanded to account for changes in sediment storage (Trimble 1977). Major changes in sediment input, transport and storage can occur because of land use, hydrological regime change or the legacy of deglaciation and volcanism. Long periods may be required for re-equilibration of the system and different portions may respond out of phase with one another (e.g. Womack and Schumm 1977; Trimble 1983; Madej and Ozaki 1996). Under these conditions, both input to and output from channel storage need to be evaluated. Evaluation methods for erosion from storage are similar to those for erosion of hillslope materials (Table 16.3); results produce estimates of alluvial sediment input due to incision or changes in channel form. Rates of aggradation on streambeds, banks and floodplains can be assessed using stratigraphic and dating methods described in previous chapters.

The catchment: integrating the sediment system

Different parts of a catchment participate in the sediment regime in different ways. Low-order channels are often the major conduits for sediment input both because they are most closely connected with hillslopes and because they account for most of the drainage density. Downstream, channels are often inset into their own deposits. These terraces and floodplains can prevent hillslope sediment from reaching the channel directly and channels at these locations may simply rework sediment initially contributed from hillslopes upstream. Opportunities for deposition and long-term storage generally increase downstream as alluvial valleys widen and gradients decrease.

The 'sediment yield' is the rate of sediment output from a catchment. Because sediment yields vary with catchment size, comparisons between catchments are usually based on yields per unit catchment area. Sediment yields per unit area frequently decrease as catchment size increases, both because average hillslope gradients decrease with increasing drainage area and because long-term aggradation is more likely downstream. Also, short-term spatial variations in precipitation and other disturbances, together with the general inverse relationship between area and disturbance intensity, create localized areas of intense erosion that can far exceed the average for the entire catchment. Transport in the channel network integrates supplies from progressively larger areas of lower erosion rate in the sampling period.

These effects have been conceptualized as creating a 'sediment delivery ratio' for a catchment, which is defined as the proportion of sediment eroded from hillslopes that is exported from the catchment. If there is a permanent sediment delivery ratio of less than 1.0, the catchment is in a state of long-term geomorphological evolution in which the high, steep uplands are being lowered relative to the lowland, which is either eroding at a slower rate or is accumulating sediment. In many catchments, however, a low sediment delivery ratio is an artefact either of the time-scale since recent disturbances or of sampling limitations. If, for example, small rainstorms during the measurement period tend to move sediment from steep areas and redeposit it on gentler slopes or in fans along a stream, a future wet period or major storm may compensate for the storage by scouring the sediment out of the catchment by water or debris flow. A wave of landscape disturbance, such as described by Haggett (1961), Trimble (1977) and many others, may also cause the lower gradient portions of the landscape to act as filters for pulses of sediment released from steeper and more disturbed areas at rates that cannot be accommodated by the catchment-scale transport system. In still other cases, a reported sediment delivery ratio seems to be a compensating artefact to correct for the fact that erosion equations such as the USLE sometimes predict unrealistically high values of sediment supply that are inconsistent with measured stream sediment transport rates or other evidence. When one is using predictive equations, it is wise to be able to identify which of these interpretations of the sediment delivery ratio is appropriate. For example, if a sediment delivery ratio is used to decrease computed sediment yields in mountainous terrain, the user should be able to explain where the sediment is coming to rest.

Early work within uniform physiographic regions of predominantly low relief (e.g. Maner 1958; Roehl 1962) showed sediment delivery ratios that decreased with increasing drainage area. The data defining these relationships were indirect estimates from the beginning. Total sediment supplies from the catchments were estimated from hillslope erosion equations and the sediment fluxes at each drainage area were measured by reservoir surveys corrected for trap efficiency. Although the original relationships have not been widely tested, they have been used elsewhere to estimate catchment sediment yield from evaluations of hillslope erosion. This is probably not an accurate prediction for many basins and thus requires some on-site confirmation. De Vente *et al.* (2007) evaluated a variety of factors that can influence downstream trends in sediment delivery ratio.

Logistical and sampling difficulties have precluded much comparison of catchment sediment yields to define their reproducibility and transferability. However, yields are generally expected to be similar for similar-sized catchments within an area of relatively uniform physiography, geology, climate, land use and vegetation cover, unless large, discrete sediment sources are present. For example, Dunne and Ongweny (1976) used average values for the forested, cultivated and grazed parts of a drainage basin, developed from a few gauged catchments, to identify major sources of sediment threatening the useful life of a reservoir. The results suggested that the original sediment yield estimate for the reservoir site was incorrect because of suspended sediment sampling limitations. In this case, new sampling surveys confirmed the calculations. Such an approach of transferring measurements from sampled to unmeasured sites requires careful consideration of differences between catchments.

The timing of sediment transport varies through a catchment. Many headwater streams cannot move clasts coarser than pebbles during frequent floods and gravel and cobbles often are trapped by woody debris. At these sites, bed material might be mobilized only when debris jams fail or during particularly large floods or debris flows. Further downstream, where bed material is finer, bed material may be mobile during ordinary bank-full events. Still further downstream, breakdown of clasts and sequestering of the larger particles lead to fining of the bed material load, until the largest rivers often transport primarily sand, silt and clay almost entirely as washload, but some fine bed material transport also occurs continuously.

16.3 Designing a sediment budget

Construction of a useful sediment budget requires the assessment of those parts of the sediment regime that are relevant to the particular application. No two applications have exactly the same goals or setting, so there is no single codifiable method for constructing sediment budgets. Sediment budgets vary widely in scope, approach and methods (Tables 16.1 and 16.4), and much of the skill of budget construction lies in deciding which form of

Table 16.4 Examples of options for sediment budget design. A particular sediment budget would be characterized by one or more options for each numbered attribute.

1. *Purpose of budget:*	5. *Temporal context:*	9. *Landscape element:*
Explain landform origin	Reconstruct past	Hillslopes
Explain change or impact	Describe present	Catchment
Describe effect of activity	Predict future	Specific landform
Describe effect of event		Altered site
Prioritize and plan remediation	6. *Duration considered:*	Channel reach
Compare systems	Event-specific	Channel system
Predict system response	Specified duration	Administrative unit
	Long-term average	
2. *Focal issue:*	Land-use activity	10. *Material:*
Landform evolution	Synthetic average	All
Land-use activity		Non-dissolved
Land-use effects	7. *Precision:*	Colluvium/soil
	Qualitative	Suspended sediment
Particular event	Order-of-magnitude	Bed material
Particular impact	Variable degree of quantification	Clay/sand/gravel
		Organic material
3. *Target of documentation or prediction:*	8. *Part of sediment regime:*	
	Weathering	11. *Method:*
Absolute amounts	Hillslope transport	Modelling
Relative amounts	Hillslope storage	Compile evidence
Description of interactions	Erosion	Inference
Locations	Delivery to channels	Analogy/transfer
Timing of response	Channel storage	Historical records
	Channel transport	Aerial photographs
4. *Spatial organization:*	Sediment attrition	Remote sensing
Distributed by sites	Sediment yield	Stratigraphic analysis
Generalized by strata	Morphological features	Monitoring
Conceptual		
Lumped		

budget is required to address the question posed. This flexibility can be a problem in applications for which the adequacy of a result is judged by whether it was obtained using standard procedures or where procedural manuals are expected to compensate for uneven levels of expertise. In such settings, it is critical to present a strong conceptual model of the sediment budget and to provide clear, well-documented explanations of the basis for the methods used. Table 16.4 lists various attributes of a sediment budget that can be selected for a particular purpose. Selection of appropriate options requires consideration of a suite of questions during design of the budgeting strategy (Table 16.5).

Identifying the study objectives

The success of a sediment budget depends strongly on the investigator's skill in defining the focal question and identifying the information needed to answer that question. To do so, the overall purpose for the inquiry must be understood. Are results of the sediment budget to be used for identifying the cause of an existing condition? To predict the outcome of future actions? To provide a basis for regulatory oversight? A single inquiry may have multiple interim objectives, but careful definition of the primary purpose allows the overall strategy to be optimized for that goal. If the budget is constructed as part of a broader project, the goals of the overall project also must be clearly articulated.

Once the primary goal is identified, it is useful to specify how sediment budget results will contribute to meeting that goal. This can be done by first identifying the kinds of conclusions or decisions to be made once results are available and then evaluating how different kinds of results might influence those outcomes. For example, if a project's ultimate goal is to reduce turbidity in a trout stream, it will be necessary to decide which sediment sources to control and which options are available to control them.

Necessary and sufficient precision

An evaluation of how sediment budget results are to be used also helps define the minimum level of precision required. If the result is intended to guide sediment control efforts, for example, a relative ranking of sediment sources based on order-of-magnitude rate estimates might be sufficient. In contrast, a study to design the management of channel sedimentation would require more precise estimates. The necessary precision can be estimated by identifying the range in potential answers over which the decisions to be supported by the study would not be altered. If the range is wide, the precision can be low.

Although many investigations would benefit from increased precision, for many others the attainable precision is higher than that actually needed to answer the relevant questions. Pursuit of unnecessary precision drains resources from other aspects of analysis where effort might be more usefully applied.

Components to be analysed

A useful sediment budget need not be complex. Most applications require exploration of only a portion of the overall sediment regime. Targeting of sediment sources for control, for example, requires assessment only of sediment input rates to channels, whereas evaluation of gravel-mining influences focuses instead on changes in sediment storage in and downstream of the affected reaches. If the intent of the budget is to determine the relative importance of a particular kind of source, it may be sufficient to evaluate the input rate from that source relative to the total sediment yield (e.g. Hovius *et al.* 1997), and budgets designed to address long-term landscape evolution generally describe the overall mass balance between sediment sources and sinks or net denudation rates rather than considering specific processes or sites (e.g. Matmon *et al.* 2003).

Identification of the portion of the sediment regime requiring study is easiest once a conceptual model has been developed for the sediment system in the area (e.g. Owens 2005). Such models have generally been in the form of flow charts (e.g. Dietrich and Dunne 1978; Reid and Dunne 1996) and tables (e.g. Kesel *et al.* 1992), but underlying each of these is the continuity equation for sediment transport, which simply states that, for some time interval, output equals input less any increase in storage. Constructing the relevant continuity equation for a particular application is useful because it requires the identification of relationships that need to be evaluated, discloses the implications of

Table 16.5 Questions useful for guiding design of sediment budgets.

Technical questions
1. What is the overall goal of the study or project of which the sediment budget is to be a part?
2. What kinds of decisions or conclusions are expected to follow from the study's results?
3. What information is needed to support those decisions or conclusions?
4. Are approaches other than sediment budgeting capable of providing that information?
5. What is the minimum level of precision needed to support the decisions or conclusions?
6. What is the minimum portion of the sediment regime that must be understood to support the decisions or conclusions?
7. To what area must the results apply?
8. To what period must the understanding apply?
Logistical questions
9. How much time is available for the study?
10. How much funding and logistical support are available?
11. What kinds of evidence are available?

disregarding particular components of the budget and identifies the information needed to balance the budget.

Different formulations of the equation are useful for different applications. Benda and Dunne (1997), for example, model the stochastic nature of sediment supply to reaches of channel throughout a network from landslides, debris flows and soil creep. To do so, they use a version of eqn. 16.1 modified to apply to individual reaches of third or higher order at specific times:

$$Q_i(k,t) + I(k,t) - Q_o(k,t) = \frac{\Delta V(k,t)}{\Delta t} \quad (16.2)$$

The terms representing input (in $m^3\ yr^{-1}$) into the channel segment k during year t are $Q_i(k,t)$, the fluvial transport (suspended and bed load) from upstream and $I(k,t)$, the sum of sediment supplied to the channel segment during the year by the processes illustrated in Fig. 16.2. $Q_o(k,t)$ is the corresponding export ($m^3\ yr^{-1}$) from segment k during year t and the final term represents the change in the volume of sediment (V, m^3) stored in segment k during year t (represented by Δt, yr).

For applications involving other kinds of information, the equation can be modified to specify the information required. For example, Dunne *et al.* (1998) examined interactions between the Amazon River and its floodplain (Fig. 16.2), so the equation used to organize the study separated the term describing storage into four parts, including deposition on bars within and adjacent to the channel (D_{bar}), diffuse overbank deposition (D_{ovrbk}), deposition in floodplain channels attached to the main channel (D_{fpc}) and deposition on the bed and banks ($A_c\rho_b\Delta z/\Delta t$, where A_c and Δz are, respectively, the area and average elevation change of the channel bed and banks in the reach, ρ_b is the bulk density of the bed material and Δt is the time interval of the computation):

$$Q_u + \sum_i Q_{trib_i} + E_{bk} = Q_d + D_{bar} + D_{ovrbk} + D_{fpc} + A_c\rho_b\frac{\Delta z}{\Delta t} + \varepsilon \quad (16.3)$$

where Q_u, Q_d and Q_{trib} are, respectively, the annual fluxes of suspended and bedload sediment at the upstream and downstream ends of each channel reach and from the i tributaries entering the reach, E_{bk} is bank erosion and ε is the error; each term has units of millions of tons per year. This equation, also, was formulated to apply to particular reaches, but in this case the results define the average annual balance of sediment transport of each grain size for each reach.

Once the underlying equation has been defined, flowcharts are useful for organizing specific information about processes. Preliminary information about major erosion and transport processes is usually available for a study area or for similar settings. This information and field inspection can be used to identify potential sediment inputs, outputs and storage changes in the area and to diagram interactions between transport processes and storage elements, with primary focus on aspects of the sediment regime on which the study is to concentrate (e.g. Fig. 16.3 and Fig. 9.6 in Chapter 9).

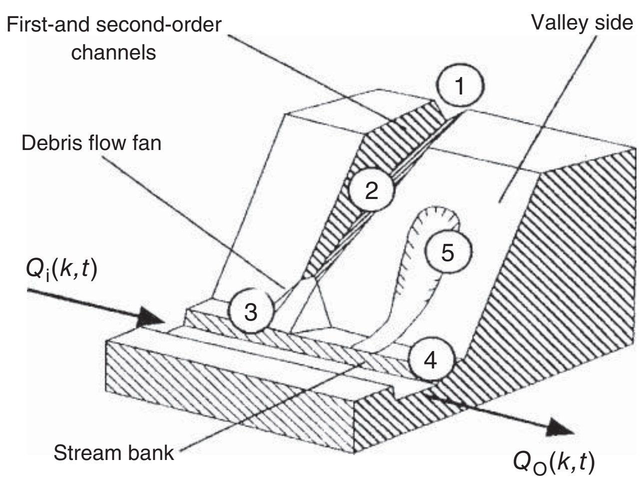


Figure 16.2 Conceptual model of the sediment budget of third- and higher order channel segments in the Oregon Coast Range. Sediment input processes include (1) shallow landsliding and debris flows in first- and second-order channels, (2) fluvial erosion and transport in first- and second-order channels, (3) bank erosion of debris flow fans and terraces, (4) soil creep along toeslopes of hillsides and (5) landslides from streamside hollows. $Q_i(k,t)$ and $Q_o(k,t)$ represent the annual fluxes of sediment load into and out of the kth segment in year t. Source: Benda and Dunne, 1997. Reproduced with permission from AGU.

Spatial scale of analysis

The most useful spatial analysis strategy for a particular study depends on the kind of area to which results are to be applied. The relevant area might be a real location (e.g. a specific catchment, channel reach or administrative district) or a hypothetical location (e.g. a 'typical' catchment or reach). If the budget is to explain conditions at a particular site, details of that site are often critical to the problem. The location of tributary inputs in a channel reach, for example, may strongly influence the functioning of the sediment budget, at least in the short term. Similarly, for budgets developed to explore landform evolution, information concerning process rates must be distributed over that landform. Geographic information systems (GIS) are useful for constructing these spatially registered sediment budgets. However, for other applications, this level of spatial specificity is unnecessary and results can be presented as averages for particular land types, land uses, sub-catchments or entire catchments. Budgets constructed for hypothetical settings, which are often used for comparing outcomes from different planning options, can be either spatially distributed or averaged.

A preliminary evaluation of the spatial distribution of processes usually simplifies budget construction by allowing the study area to be 'stratified' into areas that are likely to behave uniformly with respect to a particular process. Variables that control the rate or distribution of processes (Tables 16.2 and 16.3) provide a useful basis for stratification; these often include geological substrate and vegetation type for hillslope processes and channel order and geological substrate for channels.

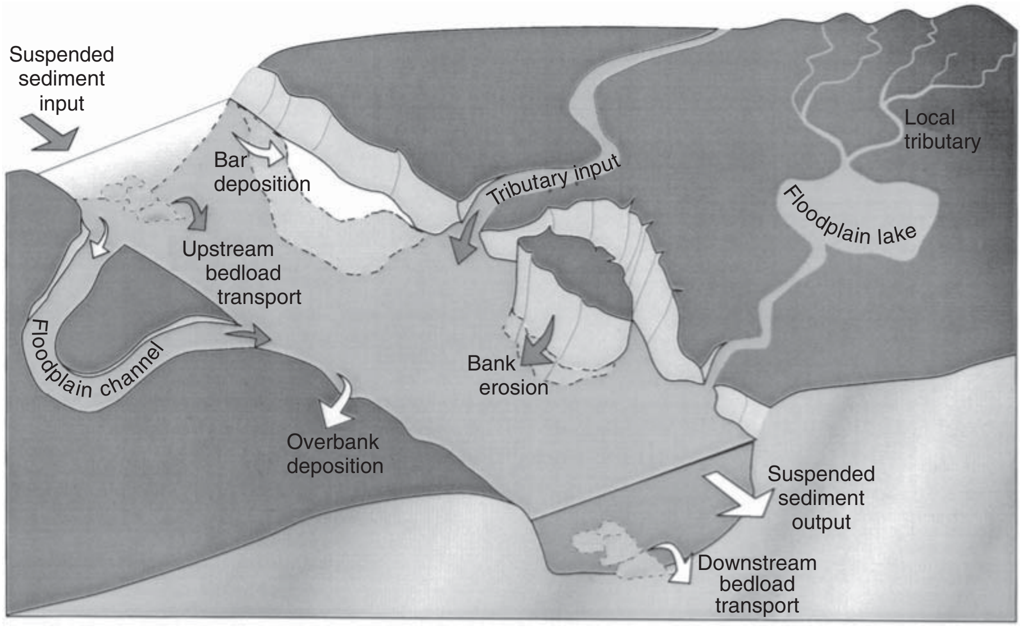


Figure 16.3 Components of the budget of channel–floodplain sediment exchanges for ∼ 200 km long reaches of the Amazon River, Brazil. Source: Dunne *et al*, 1998. Reproduced with permission of Geological Society of America.

In practice, 3–10 stratification units for each process are usually sufficient to facilitate analysis without over-simplifying the problem and a single stratification scheme often applies to multiple processes. Various methods have been used for stratification, ranging from visual delineations using aerial photographs to automated methods using GIS-based information and satellite images (e.g. Fernández *et al.* 1999; Giles 1998). For most applications, process rates and distribution are most efficiently evaluated using statistically based sampling within strata; rarely are complete inventories necessary.

Stratification allows both generalization of results across wider areas and estimation of values for particular sub-areas, and can also be used to construct budgets for hypothetical conditions. In each case, results are calculated according to the distribution of strata in the area of interest.

Temporal scale of analysis

Appropriate temporal scales can be selected for sediment budgets by considering the intended applications and the time-scales over which conditions change in the study area. A budget designed to examine the effects of land use on landsliding might evaluate landslide distribution after a single major storm on lands undergoing different uses. In contrast, a budget intended to estimate a long-term average would assess rates over a period long enough to either evaluate or average out year-to-year variations.

Budgets that evaluate changes in average sediment input relative to background conditions must consider two time-scales, the first to assess long-term average natural rates, the second to provide an analogous estimate of impacted rates. Where current conditions are changing rapidly, as is the case where land-use patterns are shifting, definition of a 'long-term average' for current conditions requires the assessment of the hypothetical response of the current land-use pattern to the long-term average distribution of triggering events. If landslide rates are defined as a function of storm size, for example, landslide incidence can be evaluated for the distribution of storms expected over a century in order to calculate the average sediment input expected if recent conditions were to be maintained for 100 years (Fig. 16.4).

Comparison of current with background conditions requires estimates of process rates active under conditions no longer present. Where nearby catchments remain in a relatively pristine condition, sediment budgets for pristine and disturbed conditions can be compared directly, but usually with significant uncertainty due to stochastic influences, such as storm occurrence. Comparisons are most feasible for catchments of similar, small size. Consequently, processes characteristic of downstream reaches often cannot be directly compared using this strategy.

More commonly, pristine examples are unavailable and catchments or hillslopes undergoing different intensities of land use are examined instead. Analysis of trends along the gradient of land-use intensities then allows inferences about pre-disturbance conditions. A gradient of hillslope conditions usually exists even where land use is uniformly distributed between catchments, so rates of hillslope processes usually can be compared even where comparison of downstream processes is not possible.

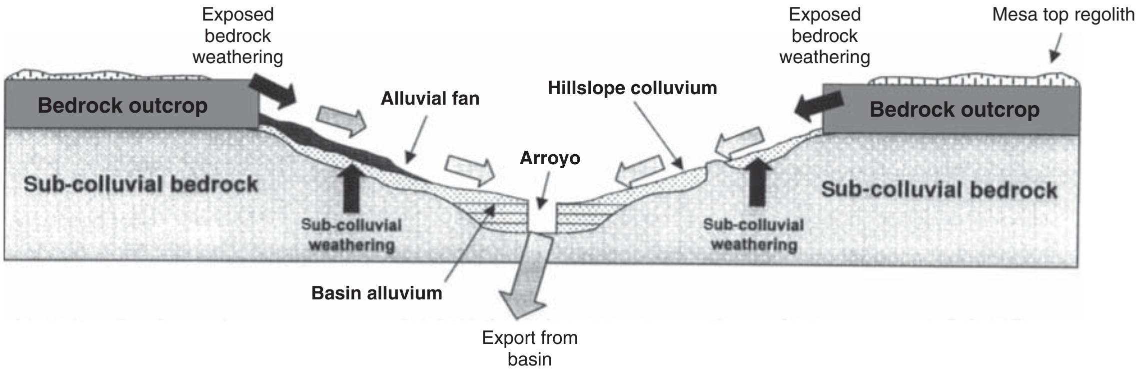


Figure 16.4 Schematic cross-section of a small catchment in New Mexico showing the general flow paths of sediment from sources (weathering of bedrock outcrops and subcolluvial bedrock) to various temporary sediment 'reservoirs' or 'stores' (hillslope colluvium, alluvial fan and valley-floor alluvium) and out of the basin. Black arrows represent mobilization by bedrock weathering. Grey arrows indicate sediment transport processes. The cross-section is approximately 1 km across and the relief is approximately 30 m. From Clapp *et al.* (2000).

Past conditions and process rates in downstream channels usually must be evaluated using evidence left by those processes or from records and accounts of earlier conditions. Stratigraphic analysis of floodplain deposits can provide considerable information about disturbance-related changes in sediment regime and channel response (e.g. Lecce and Pavlowsky 2001).

Budgets can also be designed to forecast future conditions although, as with all environmental predictions, clarity about the uncertainties involved is vital to the developer and the user of such projections. Because many erosion and sediment transport processes are strongly influenced by large, infrequent events, predictions generally describe the likely outcome, given the expected distribution of events based on a formal or informal probability analysis.

The nature of ongoing changes also influences the temporal scale appropriate for a sediment budget. If a sediment system is recovering after a major event or adjusting to a land-use change, the budget would need to incorporate a broad enough temporal scale to evaluate the nature and trajectory of the system's response.

Selection of analysis methods

Examples of analysis methods are listed in Tables 16.2 and 16.3. The methods appropriate for a particular problem depend on the nature and context of the problem, but the choice is also influenced by logistical constraints. If answers are required quickly, analysis must depend largely on existing information, sequential aerial photographs and field evidence of past process rates. If more time is available for the study, it may be useful to monitor process rates. Few studies employ only one method; different methods are used to evaluate different components of the budget or to provide multiple estimates for a single component.

Most sediment budget analyses use aerial photographs, which now provide more than 70 years of evidence of spatially distributed environmental change and are increasingly available in multispectral and digital form. Photograph sequences spanning half a century now exist for most locations and some kinds of landscape changes can be measured directly by comparison of georeferenced sequential sets. Aerial photographs are also useful for aiding landscape stratification and for planning fieldwork. Satellite imagery, now extending back for more than 40 years, can disclose alterations of land cover, changes in the position and form of large rivers and broad patterns of variation across the landscape. Old maps and survey records can also indicate changes in land use and in channel location and character. Planning departments for cities, counties and land management agencies may have GIS coverage for some attributes.

Useful information can also be provided by water quality reports, bridge surveys, flood zoning reports, reservoir surveys and stream gauging records. Trimble (2008) described the use of a variety of cultural information sources for evaluating past process rates.

Information from similar settings in other areas is also useful. In some cases, measured process rates can be transferred directly to other areas of similar character. When analogy is used to estimate process rates or interpret process interactions, similarities and differences between the study area and the measurement site need to be carefully evaluated and the justification for the transfer of information needs to be explained.

Published equations and models can be used to evaluate some process rates, but despite their availability and wide use, no model or equation can simply be assumed to be valid for a particular application. The resulting uncertainty is intensified by the general lack of validation when these models are used outside the original sites for which they were calibrated. To obtain valid results, each method's underlying assumptions, limitations and data requirements must be identified and understood and conditions for the intended application must be found to be consistent with those constraints. Wherever possible, results should be tested against those of other analytical and empirical methods or the consequent uncertainties should be highlighted as they affect the sediment budget.

Fieldwork is essential for refining the conceptual framework originally established for the budget and for checking aerial photographic interpretations. Evaluation of most chronic sediment sources requires fieldwork and fieldwork often reveals unexpected measurement opportunities. Fieldwork also allows interviews with local observers and experts at the sites of interest; general recollections can become very specific in the presence of identifiable landmarks. Fieldwork is most usefully approached both with a prioritized list of tasks to be accomplished and with an eye to finding opportunities to answer the focal questions more effectively. If possible, fieldwork should be scheduled for periods when important processes are likely to be active. Dry-season fieldwork, for example, is rarely useful for evaluating the distribution or even existence of overland flow.

Monitoring is sometimes useful during budget construction. Long-term average process rates can be defined through monitoring either if the study duration is long enough to account for temporal variations in rate (e.g. Trimble 1999) or if results define a relation between a process rate and its driving variables that allows the long-term rate to be calculated from a known distribution of driving variables (Reid and Dunne 1984; Clayton and Megahan 1986; Reid 1998). Short-term monitoring also can be useful for testing event-based modelling predictions. Comparison of modelled and monitored results for the range of sampled events indicates the level of confidence that can be placed on modelled results for unsampled events. Short-term monitoring can also reveal differences in process rates between particular site types or treatments. For any of these applications, enough sites should be monitored to provide adequate confidence that results are characteristic of the relevant site type during the monitoring period. Statistical analysis of preliminary results can identify the necessary sample size.

Integrating the results

Sediment budgets commonly incorporate disparate kinds of information and each information source usually represents a different temporal or spatial scale and a different granularity and data quality. The overall budget must reconcile these differences to produce an internally consistent, interpretable result.

Particular care must be taken to avoid mismatching timescales within a budget. Sediment budget results cannot be compared or components of a single budget combined if they represent time periods that are radically different in length or environmental conditions. For example, sediment budgets commonly incorporate monitoring data, modelling results and retrospective rate estimates. If a budget is to be checked by comparing results with 2 years of sediment yield measurements, each kind of information would need to be evaluated in such a way that results apply to that 2-year period.

Differences in spatial analysis scales are usually accounted for by stratification. A single budget, for example, might include an aerial photograph inventory of road-related landslides throughout a catchment and modelled sheet erosion rates from road surfaces on two soil types. Overall rates for both sources would vary through time as the road system developed, so inputs would be calculated per unit length of road. The average annual landslide delivery would be calculated as the total landslide delivery divided by the road-kilometre-years present during the period for which aerial photographs are available. Similarly, sheet erosion would be calculated by applying the modelled rates for each soil type to the road-kilometre-years present for that soil type during the period of aerial photographic coverage. Results could then be combined to estimate either the total input from these sources over the period of aerial photographic coverage or the combined average rate per unit length of road per year.

In general, inventory data can be used directly after suitable spatial and temporal averaging, while information characterizing particular land strata or site types is applied according to the distribution of those site types. Data that are randomly sampled without regard to site type characterize the area as a whole and cannot be used to describe portions of the sample area unless the random sampling disclosed relationships between rates and controlling variables. This pattern is also true for sampling through time: a process rate evaluated as a long-term average cannot be assumed to apply to a particular interval within the analysis period.

Auditing the sediment budget

An answer is not useful if it is not possible to determine whether it is likely to be true. Most sediment budgets represent a complex mix of calculations, mapping, measurements and qualitative inferences, so standard methods of error analysis are rarely applicable. Instead, results usually are tested by comparing estimated with measured sediment yields, assessing the reliability of each of the methods used or carrying out sensitivity analyses. The effectiveness of each approach depends on the kind of error present.

The most serious errors generally result from overlooking important processes and this can be avoided only through careful fieldwork. It is useful to begin with a complete list of major processes, identify the evidence needed to demonstrate their presence and determine whether such evidence is present. Occasionally, comparison of the summed components of the budget with a known output reveals an imbalance, but uncertainties in components are often too large for shortfalls to be revealed.

Problems have also arisen when a difficult-to-evaluate component was estimated as an unmeasured residual by subtracting the other components from a measured total. This approach implicitly assumes that all components have been identified and that the cumulative error in the sum is small enough that the difference between the sum and the total is meaningful (Kondolf and Matthews 1991). If such an approach is used, a sensitivity analysis should be carried out to identify the potential error in the result and the presence or absence of all potential budget components should be carefully verified.

Uncertainties can remain, however, even when the sediment budget is to be defined completely by empirical methods and even in reaches of river channel where one would assume that it is easier than elsewhere in the landscape to characterize and measure the various components. Both the sediment fluxes in and out of a reach and the volume of storage change are usually small residuals obtained from much larger absolute values and errors in each component are squared and summed to obtain the uncertainty in the final residual. Erwin *et al.* (2012) demonstrated the difficulty, at least for a single flood, using unusually precise measurements of bedload transport into and out of a 4 km long reach of the gravel-bed Provo River, Utah. Even considering the uncertainties in the transport rates, their measurements indicated that influx was an order of magnitude greater than efflux, so storage must have increased. When the researchers tried to locate the resulting change in storage with a combination of ground survey and remote sensing, they were unable to demonstrate rigorously that storage had in fact increased, although its most probable value was 1.6 times that derived from the flux estimates and equivalent to only 5 mm of depth change averaged over the reach. Maps of the change in storage also revealed coherent patterns. The greatest uncertainty therefore was in the average value, rather than the existence, of accumulation within the reach for the single flood. It remains to be seen how many floods would create enough storage to be statistically recognizable with the technology available for measuring such small changes in bed elevation. Topping *et al.* (2000) and Grams and Schmidt (2005) described even more challenging data limitations on larger rivers over longer periods.

Important errors also occur when decisions are founded on budget results that are mistakenly assumed to be precise and accurate. In many cases, a sensitivity analysis would have revealed the uncertainty in the budget and decisions could have been tempered to reflect that uncertainty or further work done to reduce it.

Large errors in individual budget components have occurred when modelling results were relied on without field evaluation or when short-term rates were assumed to represent long-term averages. Both of these approaches are inherently unreliable and can be identified through technical review. Where such methods are considered necessary, it is important that the associated uncertainty be evaluated and reported.

Although no single approach to testing budget results, or other empirical method, can ensure that the result is accurate, each is useful. Where estimated and measured sediment yields agree, the major components are not likely to be severely over- or underestimated, although compensating errors can occur. For any such test to be valid, clearly the estimates of sediment yield must be completely independent of analysis of budget components. Methods of 'fingerprinting' deposited or transported sediments to identify their provenance (Collins *et al.* 1998; Hill *et al.* 1998; Collins and Walling 2004) (see also Chapter 9) can be used to test portions of the overall budget.

Even technically valid sediment budgets can mislead if the question addressed by the budget is not relevant to the underlying problem. Central to formulation of a useful question is a strong understanding of what the problem to be addressed actually is. For problems associated with land-use activities, identified technical problems are often merely symptoms of underlying social, political or economic problems (Rossi 1998) and technical solutions that do not consider underlying causes will not be workable over the long term.

Assessing uncertainty

The reliability of specific methods used in budget construction usually can be assessed from the performance of a method at other sites or from other knowledge of process rates. In some cases, reliability can be expressed as a confidence interval, whereas in others only a maximum likely error can be estimated. If multiple methods are used to estimate the same budget component, discrepancies between methods indicate the maximum potential accuracy for the suite of methods used.

In some cases, formal error propagation analysis is possible for parts of a budget. The sediment budget for the Amazon River (Fig. 16.2) (Dunne *et al.* 1998), for example, was constructed to allow such analysis. Equation 16 was first simplified to

$$\mathrm{Q_u} + \sum_i \mathrm{Q_{trib_{\mathit{i}}}} - \mathrm{Q_d} = \frac{\Delta V}{\Delta t} + \varepsilon \qquad (16.4)$$

where Q_u, Q_d and Q_trib are, respectively, the annual fluxes of suspended and bedload sediment at the upstream and downstream ends of each channel reach and from the i tributaries entering the reach and $\Delta V/\Delta t$ represents the rate of change of the total sediment volume in storage in the reach. Sediment rating curves and flow duration curves available for each station on the main channel and each tributary were then used to analyse error propagation for the fluvial transport terms, allowing evaluation of the uncertainties in estimating $\Delta V/\Delta t$. The standard errors of the $\Delta V/\Delta t$ terms for individual reaches differed significantly from zero for the sand fraction in many reaches where the geomorphic and hydrological setting, and also independent estimates of the individual processes of sediment exchange, suggested that net erosion or deposition would occur. This was not generally the case for the larger silt–clay fraction, although subtle trends in net storage of this fraction did correlate with the same geomorphic and hydrological patterns. Also, the standard error of the storage estimate for the entire 2000 km floodplain reach (200 Mt yr^{-1}) differed significantly from zero for both size fractions and it agreed approximately with the storage of 500 Mt yr^{-1} estimated by quantifying each term in eqn. 16.2. However, the paucity of information available for specific storage fluxes prevented formal estimation of uncertainties for terms describing individual exchanges between channel and floodplain.

The various kinds of information used to construct a sediment budget ordinarily incorporate very different kinds and levels of uncertainty, so a standard calculation of uncertainty usually is not possible for the overall result. Instead, the sensitivity of the result to likely levels of uncertainty in the budget components can be assessed by recalculating the result for their estimated ranges of uncertainty. For some components, the uncertainty will be represented by a 95% confidence interval; for others, it may reflect a maximum likely error or complete removal of the component from consideration. Such calculations can indicate which components of the budget require the most careful analysis.

It is useful to distinguish between a result and the conclusions to which it leads. For example, a result might be a tabulation of sediment input by process, from which it is concluded that sediment control efforts should target road-surface erosion. This distinction allows the evaluation of how much a result must change before the conclusion is affected. The range in values for the result over which the conclusion would remain unchanged defines the operationally significant tolerance interval around the result (Reid and Page 2003).

16.4 Examples

To illustrate the variety of methods and strategies available for constructing sediment budgets, it is useful to examine two contrasting examples of how particular options were selected to address specific questions (Table 16.6). The first example represents a reconnaissance-level, order-of-magnitude budget, whereas the second incorporates more detailed analysis and higher levels of precision.

Evaluating sediment production from a hurricane in Hawaii

In 1992; Hurricane Iniki hit the 1325 km^2 island of Kauai, in the Hawaiian Islands. Officials wanted to know whether and where the hurricane might have increased future flood hazard by reducing channel capacity through aggradation. A sediment budget study was undertaken to determine whether significant aggradation was likely, to identify endangered areas and to suggest hazard reduction measures (Reid and Smith 1992).

The impacts of the hurricane were to be evaluated, so the budget had to be event-based (Table 16.6). Results did not need to be precise; comparison of the order of magnitude of sediment inputs to 'normal' values would be sufficient to determine the likelihood of a problem. Results needed to be spatially distributed to identify sites at risk. The necessary spatial resolution was at the scale of the 18 major catchments because vulnerable communities and structures are concentrated at catchment mouths. Calculations within each catchment could be spatially generalized.

The relevant standard of comparison for this application ordinarily would be the volume of sediment that the rivers can remove without undergoing significant change in form. If Iniki contributed more than this volume, aggradation could result. Aggradation was not a major problem before Iniki and single rainstorms with recurrence intervals of 100 years or less have triggered landsliding and sediment loading of at least an average year's sediment yield. It was therefore assumed that downstream aggradation would not occur if the hurricane had contributed less than an average year's sediment load to a river. The standard of comparison then became the average annual sediment input in years without hurricanes. Sediment yields have not been measured on Kauai, but various estimates suggest that yields range between 300 and 3000 t km^{-2} yr^{-1} and the distribution of old landslide scars suggests that years with intense storms have produced considerably higher yields.

Landslides were mapped by comparing 1:12,000 colour infrared aerial photographs taken before and after the storm. Most landslides triggered by the hurricane displaced only the soil profile and their average depth was estimated from average soil depths. The hurricane was relatively dry, so excess sheet erosion could have occurred only in areas bared by the hurricane, which were restricted to landslide scars. Trees blown down into streams carried sediment with their roots and these were mapped from a helicopter in streams wider than 5 m. The average frequency of blown-down trees was then extrapolated to smaller drainages. The volume of sediment carried by each rootwad was estimated from field observations. Sheetwash erosion rates, depths of soil removed by landslides and rootwad

Table 16.6 Examples of strategies selected for two sediment budgeting applications (see Table 16.4 for other options).

Sediment budget attribute	Hurricane Iniki budget (Reid and Smith 1992)	Clearwater road budget (Reid *et al.* 1981)
1. Purpose of budget	Prioritize, plan remediation	Prioritize, plan remediation
2. Focal issue	Particular event and impact	Particular land-use activity and impact
3. Form of results	Relative amounts	Absolute amounts
4. Spatial organization	Distributed by catchments	Hypothetical
5. Temporal context	Describe present	Describe present
6. Duration considered	Event-specific	Synthetic average
7. Precision	Order of magnitude	Precise
8. Part of regime	Erosion	Delivery to channels
9. Landscape element	Catchment	Land-use activity sites
10. Material	Non-dissolved	Non-dissolved
11. Method	Modelling, aerial photographs, existing evidence	Monitoring, aerial photographs, existing evidence

Table 16.7 An order-of-magnitude sediment budget for sediment contributed by Hurricane Iniki to catchments and hydrological zones on the island of Kauai, Hawaii. Expected annual sediment inputs are on the order of 1000 t km^{-2} yr^{-1}.

Watershed or zone	Increased sediment input from hurricane			
	Sheet erosion	Landslides	Uprooting	Total
1. Wainiha	++	+++	+	+++
2. Lumahai	++	+++	+	+++
3. Waioli	–	–	+	+
4. Hanalei	–	–	+	+
5. Kalihiwai	–	–	+	+
6. Kilauea	–	–	+	+
7. Anahola	–	+	+	++
8. Kapaa	–	+	+	+
9. Wailua	–	+	+	++
10. Hanamaulu	–	–	+	+
11. Huleia	–	–	+	+
12. Waikomo	–	–	+	+
13. Lawai	–	–	+	+
14. Wahiawa	–	–	+	+
15. Hanapepe	++	++	+	++
16. Canyon zone	–	–	+	+
17. Waimea	+	++	+	++
18. Na Pali zone	++	++	+	++

Symbols: –, <1 t km^{-2}; +, 1–10 t km^{-2}; ++, 10–100 t km^{-2}; +++, 100–1000 t km^{-2}.
Adapted from Reid and Smith (1992).

volumes were represented by likely maximum values so that results would represent the maximum potential input.

The orders of magnitude of the estimated storm inputs were then compared with expected average annual inputs (Table 16.7). Sediment inputs from Iniki were found to be potentially significant only in the two watersheds that contained large debris flows, but even there the storm-related sediment input was of the same order as an average year's input and so not likely to cause aggradation. On this basis, it was recommended that channel cross-sections be monitored periodically at potentially vulnerable locations, but that major mitigation efforts for sediment were not necessary. In all, the budget was constructed using about 10 hours of helicopter time, one day of fieldwork and one week of office work.

Prioritizing erosion control on roads in the Olympic Mountains, Washington, USA

The 375 km^2 Clearwater watershed, located in the Olympic Mountains of Washington State, is intensively managed for timber production by the Washington Department of Natural Resources. The Clearwater River is an important source of salmon and earlier work suggested that the presence of roads is associated with impairment of spawning habitat because fine-grained sediment eroded from roads accumulates in spawning gravels. Department staff needed to know the most important sources of road-related sediment in order to select effective erosion-control measures, so a sediment budget was constructed to evaluate the relative importance of road-related sediment sources in the area (Reid *et al.* 1981). Work on the budget required less than one person-year distributed over a 2-year period.

The only portion of the sediment regime that required analysis was sediment production to streams. The budget focused on fine-grained sediment, which had been identified as the major problem and sources not related to roads could be excluded. Results could be in the form of long-term average inputs and so did not need to be related to particular sites or time periods. The budget could therefore be spatially generalized. Because relative values were the major concern, only a moderate level of precision was needed.

Road-related landslide, sidecast erosion, gully and debris flow rates were evaluated for two watersheds using road construction records and three sequences of aerial photographs. Delivery ratios for these sources were assessed by measuring the volumes of sediment deposits at a selection of field sites and secondary erosion on landslide scars was estimated from root exposure and erosion pin measurements. Road-surface erosion was evaluated by sampling effluent from 10 culverts during 17 storms of various sizes and defining relations between sediment concentration and discharge for different intensities of road use. Similar measurements on a paved road segment allowed the isolation of roadcut and ditch erosion rates and these were also estimated from measurements at stakes and root exposures on roadcut surfaces. Only the erosion pins required a lengthy monitoring period and they turned out to be unnecessary because root exposure measurements provided analogous data representing a much longer effective sampling period.

Culvert hydrographs were reconstructed for unsampled storms using unit hydrographs. The discharge–concentration relationships then allowed estimation of the sediment yield for each storm over an 11-year period (Reid and Dunne 1984).

Delivery ratios for road-surface, ditch and roadcut sediment were estimated by determining the proportion of culverts that contribute flow directly to streams in the area. Sediment production from landslides, sheetwash and roadcut erosion was then calculated as an annual rate per kilometre for an average distribution of road types in a hypothetical 20 km^2 watershed, taking into account the distribution of road types and use intensities present in the area. Expected yields for any specific watershed could be calculated using the road distribution actually present. Because relations were quantified between road-surface sediment yields, storm intensities and traffic levels, average yields can be estimated for particular years and future yields can be estimated from projected use levels.

The results identify the sediment sources in most need of control: road-surface erosion and landslides each produce 10 times as much sediment as other sources (Table 16.8), while roadcut erosion is relatively unimportant, unlike conditions at some times in some other forested areas (Megahan *et al.* 1983, 1986). The results also show the importance of road use in generating sediment and quantify the impact of curtailing use during wet weather.

Table 16.8 Road-related production of sand and silt in a hypothetical 20 km^2 watershed in the Clearwater basin. Road density is 2.5 km km^{-2}, including a representative distribution of road-use intensities: 6% heavily used, 5% moderately used, 39% lightly used and 50% abandoned. Heavily used roads fall into the ‘temporary non-use’ category at night and weekends.

Source	Fine sediment production (t km^{-2} yr^{-1})
Landslides	40
Debris flows*	6.6
Gullies	0.4
Sidecast erosion	2.8
Secondary surface erosion on slide scars	12
Rills on landslide scars	3.2
Roadcut erosion†	(4.0)†
Road surface and roadcut:	
Heavy use	36
Temporary non-use	5.2
Moderate use	5.2
Light use	3.7
Abandoned	0.6
Total	*116*

*Only valley-wall erosion by debris flows is listed here; the triggering landslide is included in the landslide category.
†Roadcut erosion is included in the values for ‘Road surface and roadcut’, but is listed separately here to allow comparison.
After Reid *et al*. (1981).

16.5 Conclusions

Recent studies demonstrate the utility of the sediment budgeting approach for addressing a wide range of theoretical and applied problems in fluvial geomorphology (Table 16.1). The most efficiently constructed and effective budgets have been those designed to incorporate the kind and precision of information necessary and sufficient to address the questions posed. Only with a carefully defined focus can the appropriate options for budget construction be selected from the wide array available, and only with carefully defined objectives can the reliability of the resulting budget be assessed.

References

Aalto, R., Lauer, J.W. and Dietrich, W.E. 2008. Spatial and temporal dynamics of sediment accumulation and exchange along Strickland River floodplains (Papua New Guinea) over decadal-to-centennial timescales. *Journal of Geophysical Research* 113: F01S04.

Abernethy, B. and Rutherfurd, I.D. 1998. Where along a river’s length will vegetation most effectively stabilise stream banks? *Geomorphology* 23(1):55–75.

Allison, M.A., Kuehl, S.A., Martin, T.C. and Hassan, A. 1998. Importance of floodplain sedimentation for river sediment budgets and terrigenous input to the ocean: insights from the Brahmaputra–Jamuna River. *Geology* 26(2): 175–178.

Anderson, S.P. and Dietrich, W.E. 2001. Chemical weathering and runoff chemistry in a steep headwater catchment. *Hydrological Processes* 15: 1791–1815.

Asselman, N.E.M. 2000. Fitting and interpretation of sediment rating curves. *Journal of Hydrology* 234(3–4): 228–248.

Aufdenkampe, A.K., Mayorga, E., Raymond, P.A., Melack, J.M., Doney, S.C., Alin, S.R., Aalto, R.E. and Yoo, K. 2011. Riverine coupling of biogeochemical cycles between land, oceans and atmosphere. *Frontiers in Ecology and the Environment* 9(1): 53–60.

Auzet, A.-V. and Ambroise, B. 1996. Soil creep dynamics, soil moisture and temperature conditions on a forested slope in the granitic Vosges Mountains, France. *Earth Surface Processes and Landforms* 21(6): 531–542.

Barker, R., Dixon, L. and Hooke, J. 1997. Use of terrestrial photogrammetry for monitoring and measuring bank erosion. *Earth Surface Processes and Landforms* 22(13): 1217–1227.

Benda, L. and Dunne, T. 1997. Stochastic forcing of sediment storage and routing in channel networks. *Water Resources Research* 33(12): 2865–2880.

Beuselinck, L., Steegen, A., Govers, G., Nachtergaele, J., Takken, I. and Poesen, J. 2000. Characteristics of sediment deposits formed by intense rainfall events in small catchments in the Belgian Loam Belt. *Geomorphology* 32(1–2): 69–82.

Bishop, P., Hoey, T.B., Jansen, J.D. and Artza, I.L. 2005. Knickpoint recession rate and catchment area: the case of uplifted rivers in Eastern Scotland. *Earth Surface Processes and Landforms* 30: 767–778.

Bodoque, J.M., Díez-Herrero, A., Martín-Duque, J.F., Rubiales, J.M., Godfrey, A., Pedraza, J., Carrasco, R.M. and Sanz, M.A. 2005. Sheet erosion rates determined by using dendrogeomorphological analysis of exposed tree roots: two examples from Central Spain. *Catena* 64(1): 81–102.

Brasington, J., Langham, J. and Rumsby, B. 2003. Methodological sensitivity of morphometric estimates of coarse fluvial sediment transport. *Geomorphology* 53: 299–316.

Brooks, A.P. and Brierley, G.J. 1997. Geomorphic responses of lower Bega River to catchment disturbance, 1851–1926. *Geomorphology* 18(3–4): 291–304.

Bunte, K. and MacDonald L.H. 1999. *Scale Considerations and the Detectability of Sedimentary Cumulative Watershed Effects, Technical Bulletin 776*, Research Triangle Park, NC: National Council of the Paper Industry for Air and Stream Improvement, 327pp.

Cenderelli, D.A. and Kite, J.S. 1998. Geomorphic effects of large debris flows on channel morphology at North Fork Mountain, Eastern West Virginia. *Earth Surface Processes and Landforms* 23(1): 1–20.

Church, M. and Slaymaker, O. 1989. Disequilibrium of Holocene sediment yield in glaciated British Columbia. *Nature* 337: 452–454.

Clapp, E.M., Bierman, P.R., Schick, A.P., Lekach, J., Enzel, Y. and Caffee, M. 2000. Sediment yield exceeds sediment production in arid region drainage basins. *Geology* 28: 995–998.

Clayton, J.L. and Megahan, W.F. 1986. Erosional and chemical denudation rates in the southwestern Idaho Batholith. *Earth Surface Processes and Landforms* 11(4): 389–400.

Cole, J.J., Prairie, Y.T., Caraco, N.F., McDowell, W.H., Tranvik, L.J., Striegl, R.G., Duarte, C.M., Kortelainen, P., Downing, J.A., Middelburg, J.J. and Melack, J. 2007. Plumbing the global carbon cycle: integrating inland waters into the terrestrial carbon budget. *Ecosystems* 10: 171–184.

Collins, A.L. and Walling, D.E. 2004. Documenting catchment suspended sediment sources: problems, approaches and prospects. *Progress in Physical Geography* 28(2): 159–196.

Collins, B.D. and Dunne, T. 1986. Erosion of tephra from the 1980 eruption of Mount St. Helens. *Geological Society of America Bulletin* 97(7): 896–905.

Collins, B.D. and Dunne, T. 1989. Gravel transport, gravel harvesting and channel-bed degradation in rivers draining the southern Olympic Mountains, Washington, USA. *Environmental Geology and Water Science* 13(3): 213–224.

Collins, A.L., Walling, D.E. and Leeks, G.J.L. 1998. Use of composite fingerprints to determine the provenance of the contemporary suspended sediment load transported by rivers. *Earth Surface Processes and Landforms* 23(1): 31–52.

Cook, K.L., Whipple, K.X., Heimsath, A.M. and Hanks, T.C. 2009. Rapid incision of the Colorado River in Glen Canyon – insights from channel profiles, local incision rates and modeling of lithologic controls. *Earth Surface Processes and Landforms* 34(7): 994–1010.

Corominas, J. and Moya, J. 1999. Reconstructing recent landslide activity in relation to rainfall in the Llobregat River basin, Eastern Pyrenees, Spain. *Geomorphology* 30(1–2): 79–93.

Couper, P.R. and Maddock, I.P. 2001. Subaerial river bank erosion processes and their interaction with other bank erosion mechanisms on the River Arrow, Warwickshire, UK. *Earth Surface Processes and Landforms* 26(6): 631–646.

Davis, J., Bird, J., Finlayson, B. and Scott, R. 2000. The management of gravel extraction in alluvial rivers: a case study from the Avon River, southeastern Australia. *Physical Geography* 21(2): 133–154.

Descheemaeker, K., Nyssen, J., Rossi, J., Poesen, J., Haile, M., Raes, D., Muys, B., Moeyersons, J. and Deckers, S. 2006. Sediment deposition and pedogenesis in exclosures in the Tigray highlands, Ethiopia. *Geoderma* 132(3–4): 291–314.

De Vente, J., Poesen, J., Arabkhedri, M. and Verstraeten, G. 2007. The sediment delivery problem revisited. *Progress in Physical Geography* 31(2): 155–178.

DiBiase, R.A. and Lamb, M.P. 2013. Vegetation and wildfire controls on sediment yield in bedrock landscapes. *Geophysical Research Letters* 40: 1093–1097.

Dietrich, W.E. and Dunne, T. 1978. Sediment budget for a small catchment in mountainous terrain. *Zeitschrift für Geomorphologie, Supplement Band* 29: 191–206.

Dunne, T. 1978. 1978. Rates of chemical denudation of silicate rocks in Kenya. *Nature* 274: 244–246.

Dunne, T. 1998. Critical data requirements for prediction of erosion and sedimentation in mountain drainage basins. *Journal of the American Water Resources Association* 34(4): 795–808.

Dunne, T. and Ongweny, G.S.O. 1976. A new estimate of sediment yields in the upper Tana catchment. *The Kenya Geographer* 2: 109–126.

Dunne, T., Mertes, L.A.K., Meade, R.H., Richey, J.E. and Forsberg, B.R. 1998. Exchanges of sediment between the flood plain and the channel of the Amazon River in Brazil. *Geological Society of America Bulletin* 110(4): 450–467.

Erwin, S.O., Schmidt, J.C., Wheaton, J.M. and Wilcock, P.R. 2012. Closing a sediment budget for a reconfigured reach of the Provo River, Utah, United States. *Water Resources Research* 48(10): W10512.

Faustini, J.M. and Jones, J.A. 2003. Influence of large woody debris on channel morphology and dynamics in steep, boulder-rich mountain streams, western Cascades, Oregon. *Geomorphology* 51(1–3): 187–205.

Fernández, C.I., del Castillo, T., Hamdouni, R.E. and Montero, J.C. 1999. Verification of landslide susceptibility mapping: a case study. *Earth Surface Processes and Landforms* 24(6): 537–544.

Flemings, P.B. and Jordan, T.E. 1989. A synthetic stratigraphic model of foreland basin development. *Journal of Geophysical Research* 94(B4): 3851–3866.

Gabet, E.J. 2003. Sediment transport by dry ravel. *Journal of Geophysical Research* 108(B1): 2049.

Gabet, E.J. and Dunne, T. 2002. Landslides on coastal sage-scrub and grassland hillslopes in a severe El Nino winter: the effects of vegetation conversion on sediment delivery. *Geological Society of America Bulletin*. 114: 983–990.

Gabet, E.J., Reichman, O.J. and Seabloom, E.W. 2003. The effects of bioturbation on soil processes and sediment transport. *Annual Review of Earth and Planetary Sciences* 31: 249–273.

Gallaway, J.M., Martin, Y.E. and Johnson, E.A. 2009. Sediment transport due to tree root throw: integrating tree population dynamics, wildfire and geomorphic response. *Earth Surface Processes and Landforms* 34(9): 1255–1269.

Gao, P. 2008. Understanding watershed suspended sediment transport. *Progress in Physical Geography* 32(3): 243–263.

García-Ruiz, J.M., Lasanta, T. and Alberto, F. 1997. Soil erosion by piping in irrigated fields. *Geomorphology* 20(3–4): 269–278.

Gellis, A.C., Cheama, A. and Lalio, S.M. 2001. Developing a geomorphic approach for ranking watersheds for rehabilitation, Zuni Indian Reservation, New Mexico. *Geomorphology* 37(1–2): 105–134.

Ghimire, S.K., Higaki, D. and Bhattarai, T.P. 2006. Gully erosion in the Siwalik Hills, Nepal: estimation of sediment production from active ephemeral gullies. *Earth Surface Processes and Landforms* 31(2): 155–165.

Gilbert, G.K. 1917. *Hydraulic-mining Débris in the Sierra Nevada, US Geological Survey Professional Paper 105*, Washington, DC: US Government Printing Office, 154pp.

Giles, P.T. 1998. Geomorphological signatures: classification of aggregated slope unit objects from digital elevation and remote sensing data. *Earth Surface Processes and Landforms* 23(7): 581–594.

Giménez, R., Marazolff, I., Campo, M.A., Seeger, M., Ries, J.B., Casalí, J. and Álvarez-Mozos, J. 2009. Accuracy of high-resolution photogrammetric measurements of gullies with contrasting morphology. *Earth Surface Processes and Landforms* 34(14): 1915–1926.

Gomez, B., Eden, D.N., Hicks, D.M., Trustrum, N.A., Peacock, D.H. and Wilmshurst, J. 1999. Contribution of floodplain sequestration to the sediment budget of the Waipaoa River, New Zealand. In: Marriott, S.B. and Alexander, J., eds., *Floodplains: Interdisciplinary Approaches. Special Publication 163*, London: Geological Society of London, pp. 69–88.

Gonzalez, M.A. 2001. Recent formation of arroyos in the Little Missouri Badlands of southwestern North Dakota. *Geomorphology* 38(1–2): 63–84.

Graf, W.L. 1994. *Plutonium and the Rio Grande: Environmental Change and Contamination in the Nuclear Age*, New York: Oxford University Press, 329pp.

Grams, P.E. and Schmidt, J.C. 2005. Equilibrium or indeterminate? Where sediment budgets fail: sediment mass balance and adjustment of channel form, Green River downstream from Flaming Gorge Dam, Utah and Colorado. *Geomorphology* 71(1–2): 156–181.

Guthrie, R.H., Hockin, A., Colquhoun, L., Nagy, T., Evans, S.G. and Ayles, C. 2010. An examination of controls on debris flow mobility: evidence from coastal British Columbia. *Geomorphology* 114(4): 601–613.

Haggett, P.E. 1961. Land-use and sediment yield in an old plantation tract of the Serra do Mar, Brazil. *Geographical Journal* 127(1): 50–62.

Hall, K., Boelhouwers, J. and Driscoll, K. 1999. Animals as erosion agents in the alpine zone: some data and observations from Canada, Lesotho and Tibet. *Arctic, Antarctic and Alpine Research* 31(4): 436–446.

Harvey, A.M. 1992. Process interactions, temporal scales and the development of hillslope gully systems: Howgill Fells, northwest England. *Geomorphology* 5(3–5): 323–344.

Hassan, M.A., Church, M. and Schick, A.P. 1991. Distance of movement of coarse particles in gravel bed streams. *Water Resources Research* 27(4): 503–511.

Heimsath, A.M., Dietrich, W.E., Nishiizumi, K. and Finkel, R.C. 1997. The soil production function and landscape equilibrium. *Nature* 388: 358–361.

Hill, B.R., DeCarlo, E.H., Fuller, C.C. and Wong, M.F. 1998. Using sediment 'fingerprints' to assess sediment-budget errors, North Halawa Valley, Oahu, Hawaii, 1991–1992. *Earth Surface Processes and Landforms* 23(6): 493–508.

Hill, H. 1896. Denudation as a factor of geologic time. *Transactions and Proceedings of the New Zealand Institute* 28(11th of New Series): 666–679.

Hoffmann, T., Erkens, G., Gerlach, R., Klostermann, J. and Lang, A. 2009. Trends and controls of Holocene floodplain sedimentation in the Rhine catchment. *Catena* 77(2): 96–106.

Hovius, N., Stark, C.P. and Allen, P.A. 1997. Sediment flux from a mountain belt derived by landslide mapping. *Geology* 25(3): 231–234.

Ibsen, M.-L. and Brunsden, D. 1996. The nature, use and problems of historical archives for the temporal occurrence of landslides, with specific reference to the south coast of Britain, Ventnor, Isle of Wight. *Geomorphology* 15(3–4): 241–258.

Jackson, M. and Roering, J.J. 2009. Post-fire geomorphic response in steep, forested landscapes: Oregon Coast Range, USA. *Quaternary Science Reviews* 28(11–12): 1131–1146.

Jacobson, R.B. and Gran, K.B. 1999. Gravel sediment routing from widespread, low-intensity landscape disturbance, Current River Basin, Missouri. *Earth Surface Processes and Landforms* 24(10): 897–917.

James, L.A. 1997. Channel incision on the Lower American River, California, from streamflow gage records. *Water Resources Research* 33(3): 485–490.

Janda, R.J. 1971. An evaluation of procedures used in computing chemical denudation rates. *Geological Society of America Bulletin* 82(1): 67–80.

Kesel, R.H. 2003. Human modifications to the sediment regime of the Lower Mississippi River flood plain. *Geomorphology* 56(3–4): 325–334.

Kesel, R.H., Yodis, E.G. and McCraw, D.J. 1992. An approximation of the sediment budget of the lower Mississippi River prior to major human modification. *Earth Surface Processes and Landforms* 17(7): 711–722.

Kinnell, P.I.A. 2010. Event soil loss, runoff and the Universal Soil Loss Equation family of models: a review. *Journal of Hydrology* 385(1–4): 384–397.

Knighton, A.D. 1991. Channel bed adjustment along mine-affected rivers of northeast Tasmania. *Geomorphology* 4(3–4): 205–219.

Knox, J.C. 2006. Floodplain sedimentation in the Upper Mississippi Valley: natural versus human accelerated. *Geomorphology* 79(3–4): 286–310.

Kondolf, G.M. and Matthews, W.V.G. 1991. Unmeasured residuals in sediment budgets: a cautionary note. *Water Resources Research* 27(9): 2483–2486.

Kondolf, G.M. and Swanson, M.L. 1993. Channel adjustments to reservoir construction and gravel extraction along Stony Creek, California. *Environmental Geology* 21(4): 256–269.

Korup, O. 2005. Large landslides and their effect on sediment flux in South Westland, New Zealand. *Earth Surface Processes and Landforms* 30(3): 305–323.

Kuenen, P.H. 1956. Experimental abrasion of pebbles. 2. Rolling by current. *Journal of Geology* 64: 336–368.

Lancaster, S.T. and Casebeer, N.E. 2007. Sediment storage and evacuation in headwater valleys at the transition between debris-flow and fluvial processes. *Geology* 35(11): 1027–1030.

Lane, S.N., Tayefi, V., Reid, S.C., Yu, D. and Hardy, R.J. 2007. Interactions between sediment delivery, channel change, climate change and flood risk in a temperate upland environment. *Earth Surface Processes and Landforms* 32(3): 429–446.

Lecce, S.A. and Pavlowsky, R.T. 2001. Use of mining-contaminated sediment tracers to investigate the timing and rates of historical flood plain sedimentation. *Geomorphology* 38(1–2): 85–108.

Le Pera, E. and Sorriso-Valvo, M. 2000. Weathering, erosion and sediment composition in a high-gradient river, Calabria, Italy. *Earth Surface Processes and Landforms* 25(3): 277–292.

Lewin, J. and Brewer, P.A. 2002. Laboratory simulation of clast abrasion. *Earth Surface Processes and Landforms* 27(2): 145–164.

Liébault, F. and Piégay, H. 2001. Assessment of channel changes due to long-term bedload supply decrease, Roubion River, France. *Geomorphology* 36(3–4): 167–186.

Lisle, T.E. and Hilton, S. 1999. Fine bed material in pools of natural gravel bed channels. *Water Resources Research* 35(4): 1291–1304.

Lisle, T.E., Cui, Y., Parker, G., Pizzuto, J.E. and Dodd, A.M. 2001. The dominance of dispersion in the evolution of bed material waves in gravel-bed rivers. *Earth Surface Processes and Landforms* 26(13): 1409–1420.

Lloyd, S.D., Bishop, P. and Reinfelds, I. 1998. Shoreline erosion: a cautionary note on using small farm dams to determine catchment erosion rates. *Earth Surface Processes and Landforms* 23(10): 905–912.

Madej, M.A. and Ozaki, V. 1996. Channel response to sediment wave propagation and movement, Redwood Creek, California, USA. *Earth Surface Processes and Landforms* 21(10): 911–927.

Malamud, B.D., Turcotte, D.L., Guzzetti, F. and Reichenbach, P. 2004. Landslide inventories and their statistical properties. *Earth Surface Processes and Landforms* 29(6): 687–711.

Malmon, D.V., Dunne, T. and Reneau, S.L. 2002. Predicting the fate of sediment and pollutants in river floodplains. *Environmental Science and Technology* 36(9): 2026–2032.

Maner, S.B. 1958. Factors influencing sediment delivery rates in the Red Hills physiographic area. *EOS American Geophysical Union Transactions* 39: 669–675.

Mantovani, F., Soeters, R. and Van Westen, C.J. 1996. Remote sensing techniques for landslide studies and hazard zonation in Europe. *Geomorphology* 15(3–4): 213–225.

Marden, M., Rowan, D. and Phillips, C. 2006. Sediment sources and delivery following plantation harvesting in a weathered volcanic terrain, Coromandel Peninsula, North Island, New Zealand. *Australian Journal of Soil Research* 44(3): 219–232.

Marron, D.C. 1992. Floodplain storage of mine tailings in the Belle Fourche River system: a sediment budget approach. *Earth Surface Processes and Landforms* 17(7): 675–685.

Matisoff, G., Bonniwell, E.C. and Whiting, P.J. 2002. Radionuclides as indicators of sediment transport in agricultural watersheds that drain to Lake Erie. *Journal of Environmental Quality* 31(1): 62–72.

Matmon, A., Bierman, P.R., Larsen, J., Southworth, S., Pavich, M. and Caffee, M. 2003. Temporally and spatially uniform rates of erosion in the southern Appalachian Great Smoky Mountains. *Geology* 31(2): 155–158.

McLean, D.G. and Church, M. 1999. Sediment transport along lower Fraser River: 2. Estimates based on the long-term gravel budget. *Water Resources Research* 35(8): 2549–2559.

McLean, D.G., Church, M. and Tassone, B. 1999. Sediment transport along lower Fraser River: 1. Measurements and hydraulic computations. *Water Resources Research* 35(8): 2533–2548.

Megahan, W.F., Seyedbagheri, K.A. and Dodson, P.C. 1983. Long-term erosion on granitic roadcuts based on exposed tree roots. *Earth Surface Processes* 8(1): 19–28.

Megahan, W.F., Seyedbagheri, K.A., Mosko, T.L. and Ketcheson, G.L. 1986. Construction phase sediment budget for forest roads on granitic slopes in Idaho. In: Hadley, R., ed., *Drainage Basin Sediment Delivery, IAHS Publication No. 159*, Wallingford: International Association of Hydrological Sciences, pp. 31–39.

Merritt, W.S., Letcher, R.A. and Jakeman, A.J. 2003. A review of erosion and sediment transport models. *Environmental Modelling and Software* 18: 761–799.

Miller, J., Germanoski, D., Waltman, K., Tausch, R. and Chambers, J. 2001. Influence of late Holocene hillslope processes and landforms on modern channel dynamics in upland watersheds of central Nevada. *Geomorphology* 38(3–4): 373–391.

Moatar, F., Person, G., Meybeck, M., Coynel, A., Etcheber, H. and Crouzet, P. 2006. The influence of contrasting suspended particulate matter transport regimes on the bias and precision of flux estimates. *Science of the Total Environment* 370(2–3): 515–531.

Moody, J.A. and Martin, D.A. 2001. Initial hydrologic and geomorphic response following a wildfire in the Colorado Front Range. *Earth Surface Processes and Landforms* 26(10): 1049–1070.

Nachtergaele, J. and Poesen, J. 1999. Assessment of soil losses by ephemeral gully erosion using high-altitude (stereo) aerial photographs. *Earth Surface Processes and Landforms* 24(8): 693–706.

Nakamura, F., Sudo, T., Kameyama, S. and Jitsu, M. 1997. Influences of channelization on discharge of suspended sediment and wetland vegetation in Kushiro Marsh, northern Japan. *Geomorphology* 18(3–4): 279–289.

Nearing, M.A., Kimoto, A., Nichols, M.H. and Ritchie, J.C. 2005. Spatial patterns of soil erosion and deposition in two small, semiarid watersheds. *Journal of Geophysical Research* 110, F04020.

Nolan, K.M. and Janda, R.J. 1995. *Movement and Sediment Yield of Two Earthflows, Northwestern California, US Geological Survey Professional Paper 1454-F*, Washington, DC: US Government Printing Office, 12pp.

Norman, S.A., Schaetzl, R.J. and Small, T.W. 1995. Effects of slope angle on mass movement by tree uprooting. *Geomorphology* 14(1): 19–27.

Notebaert, B., Verstraeten, G., Rommens, T., Vanmontfort, B., Govers, G. and Poesen, J. 2009. Establishing a Holocene sediment budget for the river Dijle. *Catena* 77(2): 150–163.

Nyssen, J., Poesen, J., Veyret-Picot, M., Moeyersons, J., Haile, M., Deckers, J., Dewit, J., Naudts, J., Teka, K. and Govers, G. 2006. Assessment of gully erosion rates through interviews and measurements: a case study from northern Ethiopia. *Earth Surface Processes and Landforms* 31(2): 167–185.

Owens, P.N. 2005. Conceptual models and budgets for sediment management at the river basin scale. *Journal of Soils and Sediments* 5(4): 201–212.

Page, M.J. and Trustrum, N.A. 1997. A late Holocene lake sediment record of the erosion response to land use change in a steepland catchment, New Zealand. *Zeitschrift für Geomorphologie N.F.* 41(3): 369–392.

Page, M.J., Trustrum, N.A. and Dymond, J.R. 1994. Sediment budget to assess the geomorphic effect of a cyclonic storm, New Zealand. *Geomorphology* 9(3): 169–188.

Pelpola, C.P. and Hicken, E.J. 2004. Long-term bed load transport rate based on aerial-photo and ground penetrating radar surveys of fan-delta growth, Coast Mountains, British Columbia. *Geomorphology* 57(3–4): 169–181.

Phillips, J.D. 1986. The utility of the sediment budget concept in sediment pollution control. *Professional Geographer* 38(3): 246–252.

Phillips, J.D. 1991. Fluvial sediment budgets in the North Carolina Piedmont. *Geomorphology* 4: 231–241.

Phillips, J.D., Slattery, M.C. and Gares, P.A. 1999. Truncation and accretion of soil profiles on coastal plain croplands: implications for sediment redistribution. *Geomorphology* 28(1–2): 119–140.

Phillips, J.D., Slattery, M.C. and Musselman, Z.A. 2004. Dam-to-delta sediment inputs and storage in the lower Trinity River, Texas. *Geomorphology* 62(1–2): 17–34.

Pitlick, J. and Van Steeter, M.M. 1998. Geomorphology and endangered fish habitats of the upper Colorado River 2. Linking sediment transport to habitat maintenance. *Water Resources Research* 34(2): 303–316.

Pizzuto, J., O'Neal, M. and Stotts, S. 2010. On the retreat of forested, cohesive riverbanks. *Geomorphology* 116(3–4): 341–352.

Provansal, M., Villiet, J., Eyrolle, F., Raccasi, G., Gurriaran, R. and Antonelli, C. 2010. High-resolution evaluation of recent bank accretion rate of the managed Rhone: a case study by multi-proxy approach. *Geomorphology* 117(3–4): 287–297.

Reid, L.M. 1998. Calculation of average landslide frequency using climatic records. *Water Resources Research* 34(4): 869–877.

Reid, L.M. and Dunne, T. 1984. Sediment production from forest road surfaces. *Water Resources Research* 20(11): 1753–1761.

Reid, L.M. and Dunne, T. 1996. *Rapid Evaluation of Sediment Budgets*, Reiskirchen: Catena Verlag, 164pp.

Reid, L.M. and Page, M.J. 2003. Magnitude and frequency of landsliding in a large New Zealand catchment. *Geomorphology* 49(1–2): 71–88.

Reid, L.M. and Smith, C.W. 1992. *The Effects of Hurricane Iniki on Flood Hazard on Kauai*, Report to the Hawaii Department of Land and Natural Resources, Honolulu, Hawaii, 26pp.

Reid, L.M., Dunne, T. and Cederholm, C.J. 1981. Application of sediment budget studies to the evaluation of logging road impact. *Journal of Hydrology (New Zealand)* 20(1): 49–62.

Reneau, S.L. 2000. Stream incision and terrace development in Frijoles Canyon, Bandelier National Monument, New Mexico and the influence of lithology and climate. *Geomorphology* 32(1–2): 171–193.

Reneau, S.L. and Dietrich, W.E. 1991. Erosion rates in the southern Oregon Coast Range: evidence for an equilibrium between hillslope erosion and sediment yield. *Earth Surface Processes and Landforms* 16(4): 307–322.

Renwick, W.H., Smith, S.V., Bartley, J.D. and Buddemeier, R.W. 2005. The role of impoundments in the sediment budget of the conterminous United States. *Geomorphology* 71(1–2): 99–111.

Roehl, J.W. 1962. *Sediment Source Areas, Delivery Ratios and Influencing Morphological Factors, IAHS Publication No. 59*, Wallingford: International Association of Scientific Hydrology, pp. 202–213.

Roering, J.L., Stimely, L.L., Mackey, B.H. and Schmidt, D.A. 2009. Using DInSAR, airborne LiDAR and archival air photos to quantify landsliding and sediment transport. *Geophysical Research Letters* 36: L19402.

Rossi, G. 1998. Erosions differ. *Zeitschrift für Geomorphologie N.F*. 42(3): 297–305.

Rumsby, B. 2000. Vertical accretion rates in fluvial systems: a comparison of volumetric and depth-based estimates. *Earth Surface Processes and Landforms* 25(6): 617–631.

Santi, P.M., deWolfe, V.G., Higgins, J.D., Cannon, S.H. and Gartner, J.E. 2008. Sources of debris flow material in burned areas. *Geomorphology* 96: 310–321.

Saunders, I. and Young, A. 1983. Rates of surface processes on slopes, slope retreat and denudation. *Earth Surface Processes and Landforms* 8(5): 473–501.

Sayer, A.M., Walsh, R.P.D. and Bidin, K. 2006. Pipeflow suspended sediment dynamics and their contribution to stream sediment budgets in small rainforest catchments, Sabah, Malaysia. *Forest Ecology and Management* 224(1–2): 119–130.

Schaetzl, R.J., Johnson, D.J., Burns, S.F. and Small, T.W. 1989. Tree uprooting: review of terminology, process and environmental implications. *Canadian Journal of Forest Research* 19(1): 1–11.

Schuerch, P., Densmore, A.L., McArdell, B.W. and Molnar, P. 2006. The influence of landsliding on sediment supply and channel change in a steep mountain catchment. *Geomorphology* 78(3–4): 222–235.

Schwab, M., Rieke-Zapp, D., Schneider, H., Liniger, M. and Schlunegger, F. 2008. Landsliding and sediment flux in the Central Swiss Alps: a photogrammetric study of the Schimbrig landslide, Entlebuch. *Geomorphology* 97(3–4): 392–406.

Singer, M.B., Aalto, R., James, L.A., Kilham, N.E., Higson, J.L. and Ghoshal, S. 2013, Enduring legacy of a toxic fan via episodic redistribution of California gold mining debris. *Proceedings of the National Academy of Sciences of the United States of America* 110(46): 18436–18441.

Sloan, J., Miller, J.R. and Lancaster, N. 2001. Response and recovery of the Eel River, California and its tributaries to floods in 1955, 1964 and 1997. *Geomorphology* 36(3–4): 129–154.

Small, E.E., Anderson, R.S. and Hancock, G.S. 1999. Estimates of the rate of regolith production using ^{10}Be and ^{26}Al from an alpine hillslope. *Geomorphology* 27(1–2): 131–150.

Stock, J.D., Montgomery, D.R., Collins, B.D., Dietrich, W.E. and Sklar, L. 2005. Field measurements of incision rates following bedrock exposure: implications for process controls on the long profiles of valleys cut by rivers and debris flows. *Geological Society of America Bulletin* 117(1–2): 174–194.

Stoffel, M., Conus, D., Grichting, M.A., Lièvre, I. and Maître, G. 2008. Unraveling the patterns of late Holocene debris-flow activity on a cone in the Swiss Alps: chronology, environment and implications for the future. *Global and Planetary Change* 60(3–4): 222–234.

Stott, T. 1997. A comparison of stream bank erosion processes on forested and moorland streams in the Balquhidder catchments, central Scotland. *Earth Surface Processes and Landforms* 22(4): 383–399.

Swanson, F.J., Fredriksen, R.L. and McCorison, F.M. 1982. Material transfer in a western Oregon forested watershed. In: Edmonds, R.L., ed., *Analysis of Coniferous Forest Ecosystems in the Western United States*, Stroudsburg, PA: Hutchinson Press, pp. 233–266.

Tamene, L., Park, S.J., Dikau, R. and Vlek, P.L.G. 2006. Analysis of factors determining sediment yield variability in the highlands of northern Ethiopia. *Geomorphology* 76(1–2): 76–91.

Ten Brinke, W.B.M., Schoor, M.M., Sorber, A.M. and Berendsen, H.J.A. 1998. Overbank sand deposition in relation to transport volumes during large-magnitude floods in the Dutch sand-bed Rhine River system. *Earth Surface Processes and Landforms* 23(9): 809–824.

Topping, D.J., Rubin, D.M. and Vierra L.E. Jr., 2000. Colorado River sediment transport 1. Natural sediment supply limitation and the influence of Glen Canyon Dam. *Water Resources Research* 36(2): 515–542.

Trimble, S.W. 1974. *Man-induced Soil Erosion on the Southern Piedmont, 1700–1970*, Ankeny, IA: Soil Conservation Society of America, 180pp.

Trimble, S.W. 1977. The fallacy of stream equilibrium in contemporary denudation studies. *American Journal of Science* 277: 876–887.

Trimble, S.W. 1983. A sediment budget for Coon Creek basin in the Driftless Area, Wisconsin, 1853–1977. *American Journal of Science* 283: 454–474.

Trimble, S.W. 1993. The distributed sediment budget model and watershed management in the Paleozoic plateau of the upper Midwestern United States. *Physical Geography* 14(3): 285–303.

Trimble, S.W. 1997. Contribution of stream channel erosion to sediment yield from an urbanizing watershed. *Science* 278: 1442–1444.

Trimble, S.W. 1999. Decreased rates of alluvial storage in the Coon Creek Basin, Wisconsin, 1975–93. *Science* 285: 1244–1246.

Trimble, S.W. 2008. The use of historical data and artefacts in geomorphology. *Progress in Physical Geography* 32(1): 3–29.

Van Asch, T.J.W. and Van Genuchten, P.M.B. 1990. A comparison between theoretical and measured creep profiles of landslides. *Geomorphology* 3(1): 45–55.

Vandaele, K., Vanommeslaeghe, J., Muylaert, R. and Govers, G. 1996. Monitoring soil redistribution patterns using sequential aerial photographs. *Earth Surface Processes and Landforms* 21(4): 353–364.

Vandekerckhove, L., Poesen, J. and Govers, G. 2003. Medium-term gully headcut retreat rates in southeast Spain determined from aerial photographs and ground measurements. *Catena* 50(2–4): 329–352.

Van Oost, K., Govers, G., de Alba, S. and Quine, T.A. 2006. Tillage erosion: a review of controlling factors and implications for soil quality. *Progress in Physical Geography* 30(4): 443–466.

Van Steijn, H. 1996. Debris-flow magnitude–frequency relationships for mountainous regions of central and northwest Europe. *Geomorphology* 15(3–4): 259–273.

Vericat, D. and Batalla, R.J. 2006. Sediment transport in a large impounded river: the lower Ebro, NE Iberian Peninsula. *Geomorphology* 79(1–2): 72–92.

Verstraeten, G. and Poesen, J. 2002. Using sediment deposits in small ponds to quantify sediment yield from small catchments: possibilities and limitations. *Earth Surface Processes and Landforms* 27(13): 1425–1439.

Walling, D.E., Owens, P.N. and Leeks, G.J.L. 1998. The role of channel and floodplain storage in the suspended sediment budget of the River Ouse, Yorkshire, UK. *Geomorphology* 22(3–4): 225–242.

Walling, D.E., Owens, P.N., Carter, J., Leeks, G.J.L., Lewis, S., Meharg, A.A. and Wright, J. 2003. Storage of sediment-associated nutrients and contaminants in river channel and floodplain systems. *Applied Geochemistry* 18(2): 195–220.

Wang, J.-J., Lu, X.X., Liew, S.C. and Zhou, Y. 2009. Retrieval of suspended sediment concentrations in large turbid rivers using Landsat ETM+: an example from the Yangtze River, China. *Earth Surface Processes and Landforms* 34(8): 1082–1092.

Ward, P.A. III, and Carter, B.J. 1999. Rates of stream incision in the middle part of the Arkansas River basin based on late Tertiary to mid-Pleistocene volcanic ash. *Geomorphology* 27(3–4): 205–228.

Wathen, S.J. and Hoey, T.B. 1998. Morphological controls on the downstream passage of a sediment wave in a gravel-bed stream. *Earth Surface Processes and Landforms* 23(8): 715–730.

Wilby, R.L., Dalgleish, H.Y. and Foster, I.D.L. 1997. The impact of weather patterns on historic and contemporary catchment sediment yields. *Earth Surface Processes and Landforms* 22(4): 353–363.

Wilcock, P.R., Kondolf, G.M., Matthews, W.V.G. and Barta, A.F. 1996. Specification of sediment maintenance flows for a large gravel-bed river. *Water Resources Research* 32(9): 2911–2921.

Wilcock, P., Pitlick, J. and Cui, Y. 2009. *Sediment Transport Primer: Estimating Bed-material Transport in Gravel-bed Rivers, General Technical Report RMRS-GTR-226*, Fort Collins, CO: US Department of Agriculture, Forest Service, Rocky Mountain Research Station, 78pp.

Williams, G. P. and Wolman, M.G. 1984. *Downstream Effects of Dams on Alluvial Rivers, US Geological Survey Professional Paper 1286*, Washington, DC: US Government Printing Office.

Wohl, E.E. and Cenderelli, D.A. 2000. Sediment deposition and transport patterns following a reservoir sediment release. *Water Resources Research* 36(1): 319–333.

Womack, W.R. and Schumm, S.A. 1977. Terraces of Douglas Creek, northwestern Colorado: an example of episodic erosion. *Geology* 5(2): 72–76.

Yanda, P.Z. 2000. Use of soil horizons for assessing soil degradation and reconstructing chronology of degradation processes: the case of Mwisanga Catchment, Kondoa, central Tanzania. *Geomorphology* 34(3–4): 209–225.

Yang, S.L., Zhang, J., Zhu, J., Smith, J.P., Dai, S.B., Gao, A. and Li, P. 2005. Impact of dams on Yangtze River sediment supply to the sea and delta intertidal wetland response. *Journal of Geophysical Research* 110: F03006.

Zhang, J.H., Lobb, D.A., Li, Y. and Liu, G.C. 2004. Assessment of tillage translocation and tillage erosion by hoeing on the steep land in hilly areas of Sichuan, China. *Soil and Tillage Research* 75(2): 99–107.

Zhang, X., Phillips, C. and Pearce, A. 1991. Surface movement in an earthflow complex, Raukumara Peninsula, New Zealand. *Geomorphology* 4(3–4): 261–272.

SECTION VI

Discriminating, Simulating and Modelling Processes and Trends

CHAPTER 17

Models in fluvial geomorphology

Marco J. Van de Wiel[1], Yannick Y. Rousseau[2] and Stephen E. Darby[3]

[1]Coventry University, Coventry, UK
[2]University of Western Ontario, London, ON, Canada
[3]University of Southampton, Southampton, UK

17.1 Introduction

Models are conceptions of physical reality that can be employed to produce qualitative or quantitative predictions. Owing to the complexity of natural systems and gaps in our knowledge, the development of a model involves simplifying this physical reality to a form that fits the available resources and permits prediction of the phenomenon (ASCE Task Committee 1998). It is these models of restructured reality that are actually used and, as a general premise, there is often a discrepancy between physical reality (the 'problem') and the model. It is, therefore, incumbent on potential users to judge if in fact the model can provide predictions that are accurate enough to solve their particular problem. The primary aim of this chapter is to provide the information and guidance necessary to help users in making this judgement. The specific contents and scope of this chapter in relation to this aim are described at the end of this introduction. However, it is initially helpful to recognize the different motives of the various individuals who have formulated the fluvial geomorphological models that have been published and used by practitioners.

In fact, fluvial geomorphological modelling tools have been developed to address problems within two broad themes: pure scientific research (i.e. theory) and river management (i.e. the application of theory to solve river-related issues). Fluvial geomorphologists have traditionally focused on the study of river landforms and the processes that create them (Kirkby 1996). However, progress in understanding these areas has been hampered because rates of landform change are typically much less than the scales at which changes in natural systems are readily observable. Hence geomorphologists often use models simply as study tools to increase their understanding of the landscape. This is because models provide the basis for aggregating from the scales of observation to the scales of interest, as well as a pivotal link between the study of process and the study of landforms (Kirkby 1996). For example, many different boundaries exist in natural alluvial channels, producing distinct zones in which the morphology, hydraulic conditions and associated channel characteristics cannot be assumed to be homogeneous (Richards 1996; Lane 1998). Furthermore, it is sometimes impossible to acquire data and develop knowledge in some areas or for certain time periods, in particular for events of high magnitude and/or rare frequency. The availability of empirical data for the analysis of process-form interactions in rivers is, therefore, limited and models are increasingly being used as a supplement to these data (e.g. Darby and Thorne 1996a; Tucker and Slingerland 1997; Bradbrook *et al.* 1998; Hodskinson and Ferguson 1998; Nicholas and Walling 1998; Coulthard *et al.* 2005; Haltigin *et al.* 2007; Lane *et al.* 2007; Nicholas and Quine 2007; Van de Wiel *et al.* 2007; Luppi *et al.* 2009; Temme *et al.* 2009; Van de Wiel and Coulthard 2010).

Accelerated upland erosion and increased (or decreased) sediment loads on river channels, river regulation and channelization for flood control, navigation and hydropower have resulted in severe damage to river ecosystems, motivating a trend of river restoration to rehabilitate aquatic ecosystems (e.g. Brookes and Shields 1996; Darby and Sear 2008). Understanding these impacts and planning restoration create a demand for robust and reliable modelling tools to predict how river systems are affected by human or naturally induced environmental changes and how system components interact with human activities and management constraints.

Models are, therefore, used to provide an insight into the functioning of the natural environment, in addition to forecasting tools to underpin sustainable river management. Although the priority given to the pure and applied objectives of the discipline will differ according to the needs of diverse user groups, in the best practice there is often no conflict between them (Kirkby 1996). As Thorne (1995) puts it, there is an 'unbreakable thread' that runs between pure research and practical applications. However, although modelling tools are undoubtedly useful to a wide and diverse range of possible users, this diversity can sometimes create problems. In particular, users who are predominantly interested in practical applications may have been trained in ecology, hydrology, planning or management, but often do not have the experience to use geomorphological modelling tools with confidence. In addition, stakeholders with diverse roles can intervene in a project that incorporates

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

modelling parts. It is expected that not all of these individuals will be modelling experts, but all should be aware of the modelling aspects that are linked to their role in the project. For instance, the model user needs to ensure that the model is suitable for the project, specify the applicability and accuracy of the model and be aware of the capabilities and limitations of the modelling tool(s) employed (Refsgaard 2000). Managers must have a good knowledge of the modelling process and be able to define the requirements in terms of accuracy, code verification and model validation. Policy makers must be able to take enlightened decisions considering aspects such as the various sources of uncertainty (Beven 1996; Stewardson and Rutherfurd 2008; Wheaton *et al.* 2008) and the context to which the results apply. Consequently, there is clear potential for models to be applied inappropriately or without detailed knowledge of the strengths and weaknesses of diverse modelling approaches. There is, therefore, a need for careful communication of these strengths and weaknesses among developers, practitioners, managers and policy makers (Wilcock *et al.* 2003).

To meet this need, we provide in this chapter a systematic review of different types of modelling tools, offering guidance on their strengths, weaknesses and scope. Specifically, we review five categories of models, namely conceptual models (Section 17.2), statistical or empirical models (Section 17.3), analytical models (Section 17.4), numerical simulation models (Section 17.5) and GIS-based models (Section 17.6). Separately, we also provide a brief overview of physical models (Section 17.7). We next present an overview of the modelling process (Section 17.8), a brief overview of the applications in fluvial geomorphology (Section 17.9) and a conceptual framework (Section 17.10) with criteria for selecting the types of models appropriate for diverse user requirements, together with information required to develop an enhanced understanding of the strengths and weaknesses of specific models within each category (Table 17.1). This may help potential users in selecting between different broad modelling approaches, but reference must still be made to the guidelines detailed in Section 17.8 for the specific details of the modelling process.

Table 17.1 Summary of the characteristics, advantages and limitations of different fluvial geomorphological modelling strategies.

Model category	Typical applications	Advantages	Limitations	Model scale
Conceptual	• Reconnaissance studies • Qualitative forecasting • Qualitative postdiction	• Rapid assessment method – good for large areas and scoping studies • Relatively simple – requires few resources and minimal background data	• Requires basic training • Qualitative results only	Available across a wide range of scales (bar to catchment)
Empirical/ statistical	• Channel design • Quantitative forecasting • Quantitative postdiction • Palaeohydrology	• Simple – these models are easy to understand and use • Input data are usually readily available	• Site specific technique – care required to avoid misapplication • No information on rates of change • Requires estimate of formative discharge • Dimensionally inconsistent	Individual cross-sections representative of short river reaches
Analytical	• Channel design • Quantitative forecasting	• Improved physical basis means these models are often valid across a range of environments • Input data requirements are usually manageable	• No information on rates of change • Requires estimate of formative discharge • Models can be quite complex	Individual cross-sections representative of short river reaches
Numerical simulation	• Channel design • Quantitative forecasting	• When calibrated, valid in a wide range of environments • Provides detailed predictions of transient adjustments	• Models are very complex and require specialist training • Input data requirements very large	In theory any, but heavily constrained by data requirements
GIS-based simulation	• As associated with implemented parent model	• Implements conceptual, statistical, analytical or numerical models in spatially explicit framework	• As associated with implemented parent model	Any, but constrained by data requirements

Finally, we use a case study (Section 17.11) to show how the proposed conceptual framework can be used to ‘steer’ the direction of fluvial geomorphological modelling research applications. It should be noted that although some attention is inevitably given to the individual submodels of flow, sediment transport and bank migration processes that control the formation of river morphology, the main discussion is restricted to broader models of channel geomorphology. Detailed reviews of flow and sediment transport modelling as topics in their own right are provided elsewhere in this volume (see Chapters 8, 18 and 19).

17.2 Conceptual models

Conceptual models are an important category of tools that provide qualitative descriptions and predictions of landform and landscape evolution. A wide range of conceptual models have been developed, with applications covering the full spectrum of geographical scales from a specific river reach, up to entire landscapes (Grant *et al.* 2013). The best known example of a fluvial geomorphological conceptual model is W.M. Davis’s theory of landscape development, the geographical cycle. Davis’s theory is based on an organic analogy drawn from the Darwinian theory of evolution (Davis 1899). The theory describes how landscapes evolve as a manifestation of structure, process and time (Sack 1992). Given certain assumptions regarding structure and process, Davis believed that landscapes evolve through successive phases of youth, maturity and old age in a deterministic sequence. A key feature of this conceptual model is that landscapes are assumed to have certain diagnostic features that enable the stage of landscape evolution to be identified. Since each stage represents part of a fixed evolutionary sequence, identification of a landscape’s present stage provides the opportunity to obtain information both on its past (postdiction) and future (prediction) configuration (Sack 1992).

The simplicity with which this type of model can be applied has meant that similar models of landform evolution have been developed to predict the evolution of a wide range of fluvial landforms. Examples include the evolution of drainage networks (Glock 1931), river planform (Keller 1972; Thompson 1986; Slingerland and Smith 1998), the evolution of cross-sectional shape and longitudinal profile of incised channels (Schumm *et al.* 1984; Simon 1989) (see also Chapter 5) and the morphology of step-pool systems (Chin 1999; Caamaño *et al.* 2012), to name but a few. A characteristic feature of conceptual fluvial geomorphological models is that they rely on the technique of space-for-time substitution in their development (Paine 1985; Schumm 1991). Inherent in these models is that the distribution of characteristic landforms across space represents the passage of time exclusively. When using space-for-time substitution, it is therefore important to compare features produced by the same processes that are operating under the same physical conditions. For example, the evolution of an incised channel in alluvium can be determined by surveying cross-sections at several locations where the channel is in alluvium, but one cannot combine data or compare channels in weak alluvium with channels in resistant alluvium or bedrock and expect to find meaningful results (Schumm 1991). Hence considerable care is required in applying conceptual models based on space-for-time substitution, to ensure the modelling application replicates the conditions under which the model was developed.

Conceptual models have also been linked with tools for quantitative analysis to develop composite modelling tools that are more powerful and robust. Particularly exciting in this respect is the increasing use of geographical information systems (GIS) to store spatial data and spatial characteristics of fluvial systems (see also Section 17.6). When linked to simple logical rules derived from conceptual models, it is possible to develop quantitative models that allow for complex analyses and manipulation of data over a broad range of spatial scales. Examples of this type of composite modelling approach include linking GIS with incised channel evolution models (Simon and Downs 1995); modelling large woody debris distributions in drainage basins (Wallerstein *et al.* 1997); prediction of chinook salmon spawning habitat (Geist and Dauble 1998); modelling channel-reach morphology in mountain drainage basins (Montgomery and Buffington 1997); catchment-scale sediment transport modelling (Viney and Sivapalan 1999); channel–groundwater interaction (Wolski *et al.* 2006); and development of conceptual models to support stream restoration (Shields *et al.* 1998; Sear *et al.* 2009) and wetland rehabilitation (O’Neill *et al.* 1997).

The main advantage of conceptual models relative to other modelling approaches (Table 17.1) lies in their relative simplicity and ease of application. However, these factors are simultaneously the main limitation. Hence conceptual models provide qualitative insight into the nature of the problem but cannot be used for quantitative forecasting. It must also be remembered that conceptual models are based on the ordering of a set of empirical observations. Care must be taken, therefore, to avoid misapplication of conceptual models to fluvial systems dissimilar in character to those for which the model was originally derived. In general, conceptual models are most often used as a first step in the modelling process, enabling the model user to develop a broad understanding of the system before they attempt to apply more complex approaches geared towards quantitative forecasting. Conceptual models are also often used to identify problematic river reaches for more detailed investigation as a part of reconnaissance studies (e.g. Simon and Downs 1995).

17.3 Statistical models

Statistical models have played a major role in the analysis of fluvial systems over the last 50 years (Rhoads 1992) (see also Chapter 20). These models have been developed using functional empirical relationships between dependent morphological variables and the independent variables of sediment load and discharge. The widespread adoption of statistical models since the World War II is associated with a quantitative revolution that has now superseded the historical-descriptive paradigm, such as embodied in Davis's 'cycle of erosion', replacing this with quantitative models such as the hydraulic geometry relations (Leopold and Maddock 1953), which were simply empirical bivariate power functions (Rhoads 1992). Arguably the best use for empirical models of downstream hydraulic geometry may well be in the identification of factors relevant in the process of morphological adjustment. Attempts to reduce the degree of scatter in early hydraulic geometry models led to the adoption of the multivariate approach, incorporating additional variables such as the type and concentration of sediment load and effects of vegetation on channel morphology and extending the range of environments to gravel bed rivers. The wide availability of data collected over many years, coupled with advances in computer processing power, has contributed to the development and refinement of multivariate hydraulic geometry equations based on very large data sets including meandering and braided river channels with sand- to cobble-size beds and covering a range of flow velocities, flow depths, longitudinal slopes and Shields parameters (e.g. Lee and Julien 2006).

As analytical (Section 17.4) and numerical simulation (Section 17.5) models have become increasingly fashionable, the empirically based statistical models have been strongly criticized (Strahler 1980; Thornes and Ferguson 1981) for their lack of a solid theoretical base. Also, as with any regression study, it must be recognized that the fitted relationships of these models may not extrapolate outside the range of conditions studied and that an occasional predictor may only be significant through sampling fluke (Ferguson 1986). In empirical relationships, certain variables represent physical quantities whereas coefficients mask or group other physical quantities that are ignored or thought to have a lesser effect on the value of the independent variable. However, the contribution of neglected processes to the characteristics of the observed phenomenon may be significant in some cases. Therefore, there is a potential for models to be erroneously applied to rivers with characteristics that are not similar to those used to derive the equations.

Despite their widespread availability and use, statistical models have a number of technical and conceptual limitations (Thornes 1977). Primary amongst the conceptual difficulties is that channel changes are, by definition, transient in nature, but the various hydraulic geometry models listed above define only steady-state behaviour. A second area of difficulty lies in the undoubted existence of thresholds and discontinuities in the regression relationships involved (Thornes 1977). However, in using standard power functions, such as with the hydraulic geometry relations, it is generally assumed that these relationships are smooth and continuous. With respect to technical problems, it has been shown that the use of a common variable on both sides of the regression equation can lead to spurious correlation (Benson 1965; Brett 2004). For instance, in hydraulic geometry, flow velocity has a functional relationship with discharge, since the discharge is defined as the product of width, depth and velocity. However, determination coefficients are often calculated for this equation (and other similar equations) to represent the goodness of fit between predicted hydraulic geometries and observations (e.g. Lee and Julien 2006). Spurious correlation can significantly affect the value of correlation coefficients and coefficients of determination, hindering the understanding of the investigated phenomena (Kanaroglou 1996). Finally, traditional regression models are dimensionally inconsistent and are limited in that they only describe simple input–output relations between the states of the independent and dependent variables (Rhoads 1992).

Hydraulic geometry investigations have attempted to address some of these difficulties through the rigorous application of more sophisticated statistical and analytical techniques (Miller 1984; Rhoads 1992). These advances have established improved conceptual and technical foundations for these models. In river ecology, hydraulic conditions and certain biological processes are better represented using measures of dispersion rather than central tendency. For instance, gamma probability functions describing flow depth and velocity data points can be used to determine the characteristic hydraulic conditions of distinct habitat types (e.g. pools, glides, riffles and runs) over a range of flow stages and thus to identify suitable habitats for specific fish species (Rosenfeld *et al.* 2011). The use of more sophisticated techniques is also relevant when available data are limited. For instance, statistical models proved useful in predicting and exploring longitudinal flow variations that are gauged at few locations (Larned *et al.* 2011). Finally, the use of statistical curves allows the transfer of a model from one river channel to another, especially in cases where basic stream characteristics are considered.

In summary, statistical models, especially simple ones, tend to have less stringent data requirements, albeit with a weaker theoretical base, than their analytical and numerical counterparts (Rhoads 1992). Statistical models generally do not incorporate physical reasoning and are often dimensionally imbalanced (Hey and Heritage 1988). Critically, applications of these models are limited to the domain of the data used to estimate the model and are scale dependent (Rhoads 1992). Despite their limitations, statistical models have made a substantial contribution to our understanding of fluvial systems. Bivariate and multivariate regression models of channel and flow geometry have generated insight into relationships amongst various components of fluvial systems and such models also serve as empirical tests of theoretical models (Rhoads 1992) (see below). The simplicity and ease of application of these models have led to their widespread use in practical applications, most commonly as a preliminary step in the design of geomorphologically stable restoration reaches (e.g. Brookes and Sear 1996; Shields 1996).

17.4 Analytical models

The limitations associated with the various empirical and statistical models reviewed above has led river scientists and engineers to seek models that are based more on the physical processes involved in the establishment of channel morphology. River engineers in particular have developed models that have a more powerful predictive element than had previously been the case. It should be noted that, like the empirical models reviewed in Section 17.3, many of these analytical models are used to predict equilibrium morphology at the scale of the river cross-section, although Pizzuto (1992) has developed an analytical modelling approach linked to the scale of the watershed.

Extremal hypothesis approaches

Natural rivers have at least three degrees of freedom of adjustment in geometry: width, depth and slope (Hey 1978). Two of these appear in the hydraulic geometry equations mentioned in Section 17.3. There is, however, almost universal agreement that regime-type morphological relationships should incorporate sediment transport and alluvial friction relationships (Bettess *et al.* 1988), although by themselves these are insufficient to enable a solution for width, depth and slope, even assuming that the sediment transport and flow resistance submodels describe these processes adequately (White *et al.* 1982; Gomez and Church 1989). Extremal hypotheses have, therefore, been proposed to provide the extra relationship necessary to close the system and enable the channel morphology to be determined. An 'extremal hypothesis' is the assumption that the equilibrium channel morphology corresponds to the morphology that maximizes or minimizes the value of a specific parameter. The term 'optimality hypothesis' refers to any extremal or non-extremal hypothesis (e.g. constant Froude number, equal energy expenditure per unit channel area) that can be employed to find the optimal state of evolution of a river channel (Paik and Kumar 2010). Examples of extremal hypotheses include the minimization of energy dissipation rate (Yang *et al.* 1981), minimization of stream power (Chang 1980, 1988) or unit stream power (Yang and Song 1979) and the maximization of friction factor (Davies and Sutherland 1983) or sediment transport rate (White *et al.* 1982). Griffiths (1984) showed that these various extremal hypotheses are closely related and, under certain conditions, essentially equivalent. In addition, Huang and Nanson (2002) demonstrated that the maximum flow efficiency hypothesis (i.e. the ratio between sediment discharge and stream power) is inherent in a range of basic flow continuity, flow resistance and bedload transport relationships for alluvial rivers and produces stable channel geometries which are in agreement with observations. However, this criterion is only valid if the conditions used in the analysis are not altered and if the channel is not subject to restrictions that prevent morphological and planform adjustments to occur. In a subsequent study, Nanson and Huang (2008) demonstrated that the principle of least action governs the behaviour and equilibrium stability of alluvial rivers.

A promising analytical modelling approach, based on the extremal hypothesis that equilibrium channel morphology is associated with the maximum bed load transporting capacity (White *et al.* 1982), was developed by Millar and Quick (1993, 1998). Unlike previous investigations, Millar and Quick included a mechanistic bank stability analysis (see below) directly into the modelling approach to determine the influence of bank stability on the stable width and depth of gravel-bed rivers with non-cohesive (Millar and Quick 1993) and cohesive (Millar and Quick 1998) bank materials. The models were successfully calibrated to assess the effect of bank vegetation on bank stability (Millar and Quick 1998), which is important in designing environmentally friendly stable channels.

Verifications undertaken by the respective authors have revealed that model predictions based on extremal hypotheses provide global, if not exacting, agreement with a wide range of observations (ASCE Task Committee 1998). In an independent assessment of the predictive capabilities of extremal hypotheses, Wang *et al.* (1986) compared model predictions with empirical data from 203 sand-bed rivers and canals and 59 gravel-bed rivers and found that the various extremal hypotheses achieved a considerable degree of predictive success. Mean discrepancy ratios (*Me*) for six different extremal hypotheses ranged from 0.84 to 1.33 and from 0.74 to 1.38 for sand-bed and gravel-bed rivers, respectively. Of the hypotheses tested, the principles of minimum stream power ($Me = 1.07$) or maximum sediment concentration ($Me = 1.05$) gave the best agreement with field data. However, experimental work by Simon (1992) and Abrahams *et al.* (1994) suggested that extremal hypotheses have not yet been properly tested under a full range of imposed conditions and constraints, and Griffiths (1984) found that extremal hypotheses, when combined with conventional sediment transport and flow resistance equations, lead to regime predictions that are incompatible with observations made in flumes and natural rivers, at least in the case of straight, unconstrained alluvial reaches in morphological equilibrium. Judgement on the apparent predictive success of these methods must, therefore, be at least partially reserved.

There also have been more fundamental critiques of extremal hypothesis theories. The main criticism of analytical modelling tools based on the use of extremal hypotheses is that they simply present a method of calculating steady-state channel dimensions while not suggesting a mechanism by which this is achieved, a fact that is also true for the statistical modelling approaches (Bettess *et al.* 1988) discussed previously. Hence these hypotheses involve an essentially metaphysical method of predicting steady-state channel dimensions which offers no explanatory power (Ferguson 1986; Nanson and Huang 2008). In addition, extremal hypotheses do not consider the dynamics of environmental systems (e.g. hydrological variations and other time-variant phenomena) and the feedbacks from non-fluvial components and can, therefore, not be used to predict time-dependent evolution (Paik and Kumar 2010).

Moreover, the theoretical justifications for extremal hypotheses are also still not entirely clear (ASCE Task Committee 1998). Yang (1971) originally proposed analogies between river elevation and temperature and between potential and thermal energy to deduce his 'law of least time rate of energy expenditure' from the thermodynamic principle of minimum rate of energy production. However, this principle is valid only in the range of linear thermodynamic processes (Davy and Davies 1979), whereas energy transformations in rivers are often highly non-linear. Davies and Sutherland (1983) also attacked the theoretical basis of extremal hypotheses, on the grounds that an assumed analogy between laminar and turbulent flow used to derive the hypotheses is fundamentally unjustified. Thus, with the theoretical justification of such hypotheses unclear, application of extremal hypotheses certainly requires a clear understanding of the physical constraints presented by geological or other boundary conditions (ASCE Task Committee 1998).

An alternative approach to model equilibrium channel cross-section dimensions is tractive force modelling, which has a strong theoretical basis, since it employs the basic laws of mechanics to obtain expressions that specify the geometry of stable channel cross-sections. The basis of the approach, which was initiated in the late 1940s by the US Bureau of Reclamation (Glover and Florey 1951; Lane 1955), is to consider the magnitude of the critical tractive stress for sediment entrainment. A 'threshold' channel form is then computed that, for a pre-specified channel gradient (Carson and Griffiths 1987), can convey the flow discharge without attaining the critical stress. The various tractive force methods are, therefore, all based on various methods of solving the fluid momentum balance to obtain the local boundary shear stress, coupled with an entrainment criterion for the sediment particles that make up the channel perimeter (ASCE Task Committee 1998). See Chapter 15 for the development of these approaches.

Bank stability analyses

An entirely different type of analytical modelling in fluvial geomorphology is not concerned with the prediction of stable channel dimensions, but rather with the prediction of changes in cross-sectional channel morphology resulting from river bank erosion. This approach recognizes that the morphologies and planform of certain river types depend, at least partially, on factors that are external to the flow itself, namely soil water and riparian vegetation. The amount of soil water affects river bank stability by modifying the degree of cohesion between individual soil particles and by eroding banks during seepage. The links between fluvial processes (e.g. fluvial erosion, seepage, bank instabilities) and the time-variant contribution of these processes (and their interaction) to the occurrence of bank failures, are fundamental to the study of river evolution, but are difficult to obtain with monitoring only (Rinaldi *et al.* 2004; Luppi *et al.* 2009). Slope stability analyses include equations that allow one to obtain the risk of failure for a specific slope, to identify the most likely failure mode (e.g. translational, rotational or cantilever) and to predict the spatial extent of the slope that would be affected by a failure and the resulting cross-sectional morphology. These analytical procedures also lend themselves well to integration within numerical models (e.g. Darby *et al.* 2002, 2007; Van de Wiel and Darby 2004; Rinaldi *et al.* 2008; Luppi *et al.* 2009) (see also Section 17.5). Such models can reveal and simulate feedbacks between the included processes and produce results that are different than those obtained by non-coupled models (Darby *et al.* 2007).

Both analytical limit equilibrium models and numerical limit analysis models can be used to perform bank stability analyses (Yu *et al.* 1998). Since the former are the most popular for practical engineering and fluvial geomorphological applications (Lam and Fredlund 1993), only this category of models is discussed here. Limit equilibrium methods were developed in the mid-20th century to assess slope stability in order to design stable embankments and cuttings in unconsolidated soil materials (Fellenius 1936; Bishop 1955; Morgenstern and Price 1965; Spencer 1967). Although improvements in the accessibility to, and capacity of, computers and specialized software have contributed to reducing the time required to assess the stability of the most complex slopes, the geotechnical equations that were developed several decades ago are still in use today. Indeed, they are increasingly being employed for bank stability analyses in fluvial geomorphology investigations, as they provide a means to determine the conditions under which mass wasting will occur, in addition to the most likely mechanism and the resulting morphology.

Bank stability analyses are usually achieved in a three-step procedure. First, potential slip surfaces, i.e. interfaces upon which mass wasting occurs, are identified. Second, a factor of safety, Fs, defined as the ratio of the forces resisting the movement of a defined block of soil over the forces driving it, is computed for each of these surfaces. A failure is imminent if the driving forces surpass the forces resisting movement, i.e. if $Fs < 1$. Finally, the most likely slip surface is identified as being the one with the lowest predicted Fs value.

Several equations have been developed to calculate Fs depending on the failure mode and on the processes that are included in the analysis. For instance, Fs for a planar slip surface was given by Simon *et al.* (1991) as

$$Fs = \frac{c'L + (W_t \cos\beta - U + F_{cp}\cos i)\tan\phi'}{W_t \sin\beta - F_{cp}\sin i} \tag{17.1}$$

where c' = effective cohesion of the soil, i = angle between resultant of hydrostatic confining force and normal to failure plane, β = failure plane angle, U = hydrostatic uplift force, F_{cp} = hydrostatic confining force, W_t = failure block weight, ϕ' = effective friction angle and L = length of failure plane. Additional refinements have been made to the planar failure model to account for soil matric suction (Casagli *et al.* 1999; Rinaldi and Casagli 1999; Simon *et al.* 2000), complex bank geometries (Osman and Thorne 1988; Darby and Thorne

1996b), soil horizons and layered banks (Darby *et al.* 2000; Simon and Collison 2002) and riparian vegetation (Abernethy and Rutherfurd 2001; Simon and Collison 2002; Van de Wiel and Darby 2007; Pollen-Bankhead and Simon 2009). Alternative equations for *Fs* can be found for cantilever failures (Thorne and Tovey 1981; Langendoen and Simon 2008; Samadi *et al.* 2011) and rotational failures (Lam and Fredlund 1993; Rinaldi *et al.* 2004). In most studies, the failure mode is imposed by the selected model or analytical solution and thus the occurrence of a specific mode is tested. However, some models account for multiple failure modes (e.g. Dapporto *et al.* 2001; Langendoen and Simon 2008).

The natural riparian environment usually includes vegetation, which modifies the geotechnical and hydrological properties of the bank (Abernethy and Rutherford 2000, 2001; Simon and Collison 2002; Pollen *et al.* 2004; Pollen-Bankhead and Simon 2010), thus affecting the bank's susceptibility to mass failure. Modelling experiments can help identify the most important effects and the conditions under which these are significant. For example, bank stability models including vegetation dynamics were used to test the hypothesis that a tree stand established near an incised stream can reduce bank erosion rates (Van de Wiel and Darby 2007; Langendoen *et al.* 2009; Pollen-Bankhead and Simon 2009). A key factor in this is the model's representation of roots, which mechanically reinforce the soil by making it stronger in tension and by reducing scour. Root characteristics (diameter, density and tensile strength) can be integrated in bank stability analyses to estimate the amount of shear strength added by roots (e.g. Van de Wiel and Darby 2007; Langendoen *et al.* 2009; Pollen-Bankhead and Simon 2009). However, these root characteristics typically are dependent on species, age and local growing conditions. Several field and laboratory experiments were conducted to estimate the tensile strength of roots of different types, age, species and regions (e.g. Simon and Collison 2002; Docker and Hubble 2008), and other studies have attempted to understand the spatial and temporal variations in overall soil reinforcement for specific species (Abernethy and Rutherfurd 2001; Pollen 2007). The architectural features of roots include topological (i.e. branching patterns) and geometric (i.e. shape, size, orientation and location of components) parameters (Reubens *et al.* 2007) and, when combined with the physical properties (i.e. tensile strength, flexibility), these explain the diversity of reinforcement values measured for different plant species. Hydrological effects of vegetation can also be integrated into bank stability analyses. Although they are usually less important than the mechanical effects, inter-species differences can be significant (Simon *et al.* 2006). Likewise, evaporation can have significant seasonally variable effects on soil moisture content and on soil cohesion (Pollen-Bankhead and Simon 2010). The greatest problem with including vegetation in bank stability analysis for applied problems lies with finding suitable parameter values to represent the vegetation adequately – not with the limitations of the models themselves. In practice, therefore, most bank stability analyses simplify the real world either by using estimated vegetation parameters, by implicitly including vegetation effects in the calibration of other model parameters or by ignoring vegetation altogether.

The general aim of bank stability analyses is to predict, with sufficient accuracy, the occurrence (i.e. timing), mode (planar, rotational, cantilever) and magnitude (volume) of a bank failure event. Some studies have shown notable predictive success. For example, Langendoen and Simon (2008) reported good agreement between simulated and real rates of bank retreat in a composite river bank, with on average 7% overprediction for the central section of the bendway and 15% underprediction at bank top. Midgley *et al.* (2012) correctly predicted timing of bank failures, but noted that the amount of bank retreat was systematically under predicted. Rinaldi *et al.* (2004) successfully predicted the timing and spatial extent of a rotational slide, and Luppi *et al.* (2009), using an analytical bank stability model in conjunction with a numerical flow model, successfully simulated the timing and magnitude of observed cantilever and rotational failures in a sequence of flow events.

Despite the success in predicting various characteristics of failure events, the uncertainty of the predictions remains problematic. Samadi *et al.* (2009) suggest that the likelihood of generating unreliable predictions due to natural variability of input parameters can be greater than 80% for a required precision lower than 15%. The description of the observed conditions is also critical. The accuracy of the prediction achieved also depends on the quality of data and on the processes included in the model. Care is recommended in sampling field conditions in order to perform reliable predictions (Samadi *et al.* 2009). For example, in the case of shear and beam cantilever failure types, bank stability is especially sensitive to variations in the geometric shape of the overhanging block and to the cohesion and unit weight of the bank material (Samadi *et al.* 2011). Luppi *et al.* (2009) suggest incorporating fluvial erosion processes to avoid overpredicting *Fs*. Incorporating vegetation might also improve accuracy, but is subject to large uncertainties in vegetation parameterization. Most field measurement techniques for obtaining root data are invasive and damaging to the vegetation. Moreover, laboratory data are not always compatible with field data. For example, tensile strength and breaking force of roots are systematically higher in the laboratory than in the field (Tosi 2007).

17.5 Numerical models

Numerical models differ from their conceptual, empirical and analytical counterparts in that they are spatially and temporally multidimensional, thereby facilitating a physically based description of fluvial processes. Advances in computational hardware capacity and the increasing number and variety of available computational fluid dynamics (CFD) software packages have contributed to the rise of two- (2D) and three-dimensional (3D) models, which are unlocking the real potential of numerical modelling for use in fluvial

Table 17.2 Representation of the real world in a numerical simulation model,

Real world	Model representation
Space dimensions	Grid (discretization)
Time dimension	Time steps (discretization)
Physical properties	Discrete values on grid
Physical processes	Governing equations
Evolution	Numerical algorithm solves equations and changes values on grid

geomorphology, i.e. the simulation of geomorphological changes over space and time. Numerical models thus allow better understanding and simulation of key fluvial processes (Lane 1998), some of which would be more difficult or impossible to study using non-numerical models. For instance, the helical motion of the flow in a meander belt, the associated inward sediment transport and the resulting creation of distinct geomorphic features (e.g. point bars) can only be simulated using numerical models. Since sediment transport and channel morphological adjustment processes depend greatly on flow characteristics, multidimensional numerical models constitute an additional opportunity to study process–form relationships in the fluvial environment, albeit not the only one (Hervouet and Van Haren 1996).

Not surprisingly, a variety of numerical models have been developed for applications in fluvial geomorphology. However, despite the differences between these models, a set of generic concepts common to all numerical models can be identified.

Concepts of numerical modelling

In a numerical model, physical space is represented by a grid or mesh consisting of a finite number of points. Spatial physical properties or characteristics (e.g. landform elevation, water depth, roughness, flow velocity) are represented on this grid by a set of discrete values. Representation of the physical processes relevant to a particular problem is achieved in two steps. First, the relevant processes are identified and described in mathematical form (i.e. a set of governing process equations is formulated). Second, a numerical algorithm is developed in order to solve or approximate the governing equations over the discretized grid. The time dimension is also discretized into time steps and temporal change or evolution is represented by changes in the values on the grid (Table 17.2).

Spatially, geomorphological processes can be viewed in a complex hierarchical context: every geomorphic system consists of a series of ever smaller, lower level systems, but is at the same time part of a sequence of ever larger, higher level systems. Depending on the scale of the system and the objective of the investigation, certain levels will be dominant whereas others play a secondary role and can be ignored (De Boer 1992). Numerical simulation of geomorphological processes implicitly involves three levels of the spatial hierarchy. The largest of these is the area under investigation, which is represented by a 2D or 3D grid. Generally, none of the geomorphological processes on this level are explicitly incorporated in the model, as prediction and study of these processes are usually the purpose of the model. The second level is represented by the individual grid element or cell. The cell forms the core of the numerical model as this is the level on which the processes are explicitly modelled. The third and smallest level of processes is commonly referred to as the subgrid-scale level. Subgrid-scale processes are modelled implicitly by aggregating their effects on the grid element level. Usually this requires assumptions concerning the spatial and temporal occurrence of subgrid-scale processes (e.g. turbulence) to be made (see Table 17.4). Subgrid-scale processes are treated as such because their explicit modelling would be too demanding on computational resources or because they are not sufficiently understood.

Numerical grids can differ in many ways: number of dimensions (one, two, three), shape of the elements (triangular, quadragonal, hexagonal), coordinate system (Cartesian, cylindrical, curvilinear). These grid attributes are generally determined by the numerical techniques used for solving the mathematical equations and, given a certain model, cannot be influenced by the user. However, the user is usually faced with constructing a grid that captures the physical world. The resolution of the grid affects the internal working of the model and its output (Olsen and Kjellesvig 1998; Hardy *et al.* 1999). It is often thought that the accuracy of model prediction increases with increasing grid resolution. This hypothesis is powered by the idea that a finer grid results in improved representation of the physical world and improved stability of the numerical algorithm (Hardy *et al.* 1999). According to this reasoning, very high predictive accuracy can be achieved only if the necessary computational resources are invested. However, this hypothesis only holds true up to a certain level, i.e. there is a limit to grid refining, beyond which further increases in spatial resolution will not result in a significant improvement of predictive accuracy (Farajalla and Vieux 1995; Bates *et al.* 1996; Hardy *et al.* 1999). In terms of the hierarchical levels, it can be said that this limit is reached when the grid resolution captures all the essential characteristics and variability of the explicitly modelled processes on the grid element level. At that stage, further improvement can only be made by explicitly modelling the subgrid-scale processes, that is, by ‘upgrading’ them to the grid element level. However, this is often computationally unachievable, as has been shown for the explicit modelling of turbulence (Hervouet and Van Haren 1996; Lane 1998).

In numerical models, the temporal dimension is discretized in time steps. The temporal scale of processes to be simulated exercises considerable control over the structure of the model and the spatial grid. Simulations over long periods, i.e. hundreds of years, usually require large time steps to be computationally efficient, which reflects back into the choice of grid resolution and the notion of what processes can be explicitly modelled on the grid element level and what should be aggregated on the subgrid scale.

Translating the relevant physical processes into a set of governing equations that can be solved by a numerical algorithm is the key element in numerical modelling. Mathematical descriptions of physical processes can either be theoretical (e.g. the equations of fluid motion) or empirical (e.g. most sediment transport equations) (Kirkby 1996). A combination of both descriptions is often required when developing models for applications in fluvial geomorphology as the computation of sediment transport rates within a channel necessitates information on fluid motion. Subsequent solution of the governing equations by means of a numerical algorithm can be achieved in numerous ways. Exactly which processes are realized, and how this is done, depend very much on the problem under investigation. However, it is important to realize that the predictive capability of a model is largely influenced by the adequacy of the descriptions of the physical processes and by the techniques used for solving the equations. These techniques, and in particular the choice of a specific algorithm, in turn affect computation time. Depending on the level of detail in the process representation, two main types of numerical model can be considered: reductionist models and reduced complexity models.

Reductionist models

The reductionist approach to modelling attempts to replicate the governing processes in as much detail as possible, with subgrid parameterization kept to a minimum. This is particularly challenging for the representation of fluid motion. The basic laws governing the motion of fluids are the 'conservation of mass' and 'conservation of momentum' (see also Chapter 18). Mathematically, these laws are expressed by a set of non-linear differential equations (Tritton 1988):

$$\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho u) = 0 \qquad (17.2)$$

$$\rho \frac{\partial u}{\partial t} = -\nabla p + \mu \nabla^2 u + F \qquad (17.3)$$

where u = velocity vector ($\mathrm{m\,s^{-1}}$), t = time coordinate (s), ρ = fluid density ($\mathrm{kg\,m^{-3}}$), p = pressure (Pa), μ = viscosity coefficient ($\mathrm{m^2\,s^{-1}}$) and F = external forces (Pa). In their general form, applying to all fluid motion, these equations are called the Navier–Stokes equations. Theoretically, these equations can be solved exactly, in all three spatial dimensions and accounting for all turbulence effects, only if the resolution of the grid and the time step of the calculations are fine enough. However, even for the simplest problems, the required resolution would lead to a very large number of grid points and unacceptable computing times (Hervouet and Van Haren 1996; Lane 1998).

Nearly all fluvial geomorphological numerical models rely on solving the Navier–Stokes flow equations in one form or another (see Chapter 18 for details). The flow equations are usually solved using either finite difference or finite element techniques (Sewell 1988), finite volume techniques (Bradbrook *et al.* 2001) or spectral methods (Olsen and Stokseth 1995; Pinelli *et al.* 1997). In a finite difference technique, the region of interest is covered by a rectangular grid and spatial derivatives are approximated by finite differences between adjacent grid elements (Vreugdenhil 1994). The benefits of using this type of technique are that the solution is easier to implement, design and understand, it requires fewer data, it is mathematically simpler and it facilitates data input (Kresic 2007). Finite element techniques offer enhanced flexibility in the representation of complex, irregular geometries and boundaries (Vreugdenhil 1994; Kresic 2007). The reduced number of grid cells can decrease computing time. In finite volume methods, the governing equations of fluids are discretized in the integral form (Kolditz 2002). Such methods possess several advantages, such as handling both structured (i.e. meshes which consist of two sets of lines) and unstructured meshes, conservating quantities at a local scale, easily introducting natural boundary conditions and approximating complex geometries (Kolditz 2002). Finally, spectral multidomain methods approximate a numerical solution of partial differential equations as a linear combination of continuous functions (Gervasio *et al.* 1997). These methods, when used in combination with the projection decomposition method, are accurate, involve low phase error, are compatible with complex geometries and are suitable for approximation with parallel computing (Pinelli *et al.* 1997). The accuracy achieved, however, varies with the degree of smoothness of the solution (Gervasio *et al.* 1997). Solution techniques differ mainly in their ease of implementation, computational efficiency and conservation properties, although they can also cause minor variations in model output (Bates *et al.* 1996). In general, however, finite difference techniques are easier to implement, whereas finite element and finite volume techniques use less computing time, achieve higher order accuracy and allow more freedom in assembling the computational grid, which is convenient when dealing with highly irregular geometries, such as the morphology of natural river channels (Rice 1983; Sewell 1988). Without considering sediment transport, these CFD models can be used to investigate hydraulic flow properties over an unchanging topography. Examples include the simulation of flood propagation and inundation extent (e.g. Beffa and Connell 2001; Horritt and Bates 2001, 2002; Dutta *et al.* 2007), of flow patterns around bedforms and in-channel engineering structures (e.g. Jia *et al.* 2005; Haltigin *et al.* 2007), of dam break scenarios (e.g. Hervouet 2000; Liao *et al.* 2007) or of habitat suitability for aquatic species (e.g. Booker and Dunbar 2004; Clifford *et al.* 2008; Daraio *et al.* 2010). These type of applications are mainly used in planning or river engineering studies.

For more strict geomorphological application, CFD models need to be coupled with sediment transport equations (see also Chapter 18). The calculation of sediment entrainment, transport and deposition allows for the simulation of changes to the topography and morphology of the modelled river system. These type of models are more complex, partly because they need to consider the sediment transport algorithms, but mainly because the changes to the topography require the underlying topographic grid to be updated periodically. Such changes may have feedbacks on the flow field and hence require the flow field to be recalculated after every change to the topographic grid, which is computationally very demanding. Applications of these coupled models are, therefore, constrained to relatively small spatial and temporal scales, typically focusing on smaller channel sections, such as individual bends or confluences (e.g. Kassem and Chaudhry 2002; Ferguson *et al.* 2003; Rüther and Olsen 2005; Fischer-Antze *et al.* 2009; Mekonnen *et al.* 2010) or on relatively short channel reaches (e.g. Fang and Wang 2000; Guo and Jin 2002). These models mainly focus on simulating the evolution of channel bed features such as pools and riffles (e.g. Booker *et al.* 2001), bars (e.g. Nicholas and Sambrook Smith 1999), dunes (e.g. Maddux *et al.* 2003; Stoesser *et al.* 2008) or concave bank benches (e.g. Vietz *et al.* 2006).

The models can be further expanded to model the wider fluvial context also, for example by modelling morphological change of the floodplain through overbank sediment routing and deposition (e.g. Nicholas and Walling 1998; Hardy *et al.* 2000; Thonon *et al.* 2007) or by modelling planform channel changes due to bank erosion or width adjustment (e.g. Nagata *et al.* 2000; Darby *et al.* 2002; Duan and Julien 2005; Crosato and Saleh 2010) (see also Section 17.11).

The reductionist models thus exhibit increasing levels of complexity depending on how many features are being modelled. Starting with strict flow hydraulics models, additional functionality can be gained by adding more processes, for example, the simulation of sediment transport, lateral erosion or interaction with riparian vegetation (Van de Wiel and Darby 2004). However, these increases in functionality come with corresponding increases in model complexity. Furthermore, the accuracy of the simulations typically decreases with increasing complexity as the number of model variables and parameters increases from one group to the next owing to the incorporation of additional processes, each with their own range of measurement error or uncertainty. For instance, flow patterns in a river channel can be reasonably well predicted by a 3D CFD model, but accurate quantitative spatial predictions of planform changes are more challenging as the number of influencial variables can be significant (e.g. bed and bank geology, hydrology, topography and land cover, rather than strictly flow hydraulic variables).

Reduced complexity models

Reductionist numerical models can quickly become very complex as ever more physical processes are included, at ever smaller scales. To counter this trend towards increased detail and complexity in model structure, an alternative and more holistic paradigm of numerical modelling has developed over the last decade, based on the observation from chaos theory that simple rules can lead to complex dynamic behaviour. These models, commonly termed 'reduced complexity models' (Brasington and Richards 2007; Coulthard *et al.* 2007; Murray 2007; Nicholas and Quine 2007; Nicholas 2010) (RCMs), simplify the representation of the physics of the governing processes, instead relying on simple and often empirical equations. For example, instead of approximating the Navier–Stokes equations, many RCMs use Manning's equation to calculate the flow field. In using this equation, steady or quasi-steady flow is assumed and flow calculations are essentially driven by topography and conservation of mass, ignoring conservation of momentum or secondary flow circulation. Effectively, RCMs attempt to capture the essential physics (Brasington and Richards 2007), i.e. the simplest representation of processes that still produces acceptable results (Murray 2007). This is in contrast to reductionist modelling where ever more detail is added to increase model realism, as long as computational efficiency and model uncertainty permit. Note, however, that the RCMs are still process based in concept, albeit that they simulate the effects of the processes rather than the processes themselves.

RCMs often are cellular models, i.e. raster-based models where local rules define interactions between neighbouring cells (e.g. fluxes of water and sediment as a function of flow depth, slope and grain size). One of the first models in fluvial geomorphology to rely on this approach was the braided river model of Murray and Paola (1994, 1997), which coupled the topographic flow routing rule with a simple sediment transport rule. By repeatedly iterating these rules, the model could simulate seemingly realistic braid patterns and planform dynamics of braided rivers (Murray and Paola 1994, 1997). Since then, several RCMs have been developed for simulation in fluvial geomorphology, including models for bank erosion (Fonstad and Marcus 2003; Coulthard and Van de Wiel 2006), bar migration (Nicholas 2010), channel flow (Nicholas 2009), vegetation effects in braided rivers (Murray and Paola 2003; Coulthard *et al.* 2007), alluvial fan building (Coulthard *et al.* 2002) and catchment-scale landscape evolution (Willgoose *et al.* 1991; Tucker and Slingerland 1997; Coulthard *et al.* 2005; Leyland and Darby 2009; Temme *et al.* 2009).

RCMs, like reductionist models, rely on behaviour on the system scale emerging from the rules defined at the grid scale. This emergence of patterns and dynamics at a larger scale is a characteristic of all numerical models. Indeed, it is an essential property, since without it nothing could be learned from either reductionist models or RCMs (Van de Wiel *et al.* 2007). However, the nature of the emergence is different between the two approaches, which influences the types of questions asked (Bras *et al.* 2003; Murray 2003, 2007). Reductionist models are predominantly used for detailed predictive modelling, i.e. for making specific predictions about systems where the processes and their interactions are well understood. RCMs, on the other hand, are mostly used for exploratory studies, i.e. for systems where the processes or their interactions are not typically not well understood and where the simulations may generate new insights. The advantages of RCMs for this type of modelling are twofold. First, the models rely on simple, often intuitive, rules which typically are easy to understand and implement. Second, and more important, the relative simplicity of the rules facilitates the analysis of the results and permits tracing the origins of emergent dynamics (e.g. Van de Wiel and Coulthard 2010).

The strength of RCMs is also their disadvantage. Because they only simulate the essential physics, rather than all the influencing factors, they cannot make detailed predictions in space and time. For example, although the braided river model from Murray and Paola (1994, 1997) can simulate realistic braiding patterns and dynamics, including braiding index, channel avulsion and lateral migration, it cannot predict the morphological changes at a specific point in any specific braided river. Other RCMs suffer from similar shortcomings. This does not render RCMs useless, but it does mean that researchers have to be careful how simulation results are interpreted.

Finally, it should be noted that RCMs and fully reductionist models are the extreme endpoints of a continuous scale (Murray 2003; Werner 2003). In practice, most numerical models are neither fully RCM nor fully reductionist; instead, they operate somewhere in between, with computational fluid dynamics tending towards the reductionist end of the scale and cellular models tending towards the RCM end.

Benefits and disadvantages

The main overall benefit of numerical models, both reductionist and RCMs, is that they can simulate geomorphological change over space and time, i.e. they can simulate the *evolution* of the system being modelled. This permits a comprehensive analysis of the geomorphological development of a given area and is therefore a valuable tool for generating insights and understanding. Moreover, the ability to perform repeated simulations, thus allowing sensitivity tests and what-if scenarios, can be a valuable aid in environmental planning and land or resource management. Finally, numerical models provide a vast amount of data as output (often continuous over the simulation domain), which can easily be summarized, analysed or visualized.

However, numerical modelling is also inherently dependent on an adequate representation of the physical world. Accurately translating physical processes into an acceptable set of governing equations and elaborating an efficient solution algorithm can be challenging, and obtaining the vast amount of necessary field data can be difficult and expensive. In addition, numerical models can be very demanding on computer resources, depending on the construction of the model, the duration and complexity of the simulated processes and the required level of accuracy. Consequently, numerical models typically produce very detailed output data, which can be overwhelming and which can inspire unjustified faith in their accuracy. The fact that models are often accompanied by user-friendly graphical interfaces that hide implemented rules and algorithms enhances the risk for their misuse (Hervouet and Van Haren 1996). For this reason, it is important that the user, when interpreting modelling results, is aware of the underlying assumptions and inherent limitations of achieved predictions. In many cases, these disadvantages are outweighed by the benefits. In addition, numerical models may be the single category of model that can address certain questions.

17.6 Use of remote sensing and GIS in fluvial geomorphological modelling

Modelling in fluvial geomorphology, particularly statistical modelling and numerical modelling, require the acquisition and manipulation of large amounts of data. The use of GIS and remote sensing tools has become almost unavoidable.

Remote sensing, particularly airborne remote sensing, permits the rapid acquisition of high-resolution continuous data over large areas (see also Chapter 6). In contrast, the use of many ground-based surveying techniques is more time consuming and can disturb the studied sites (Couper *et al.* 2002), and other rudimentary data sources (e.g. topographic maps) are often not accurate enough to detect landforms (Notebaert *et al.* 2009). Remotely sensed data are especially suitable for use with numerical and statistical models which typically require large amounts of data. These data may result from the interpretation of aerial photographs, radar imagery or light detection and ranging (LiDAR). For instance, the analysis of airborne or space-borne photographs allows the delineation of river-floodplain boundaries, the quantification of the migration rate of rivers (Yao *et al.* 2011), the identification of the geometrical properties of a channel and the estimation of flow discharge in a large river based on its width (Brakenridge *et al.* 2002). In addition to airborne LiDAR, ground-based LiDAR systems were found to be useful in examining the effects of vegetation on bank morphology evolution due to their capacity to detect bank undercutting and erosion between adjacent trees (Pizzuto *et al.* 2010).The characteristics of groundcover can also be derived from the analysis of spectral bands from radar imagery. Finally, the analysis of LiDAR data and stereoscopic pairs (De Rose and Basher 2011) can be used to create digital elevation models (DEM) of the channel surface and the near-channel topography.

A geographical information system mainly provides a framework for the integration and standardization of spatial data. In particular, modelling results can be stored, manipulated, analysed and visualized (Downs and Priestnall 1999). The set of tools provided by software developers and the contribution of the community of GIS users make numerical, statistical and analytical models more explicit then they would otherwise be. Flow routing, interpolation and conversion tools are especially useful in the context of river-related analysis.

There are, however, some limitations regarding potential uses of GIS and remote sensing in fluvial geomorphology. Whereas remote sensing is usually referred to as providing high accuracy and high spatial resolution data, the characteristics of the acquired data and the amount of effort required for their interpretation vary with the technology used. GIS can potentially introduce uncertainties associated with the structure with which the computer stores data. For example, analysing the topography of a small river channel using GIS may be inappropriate unless the spatial resolution is high enough to allow the identification of the landforms or channel features of interest (Notebaert *et al.* 2009) or unless topography is represented in vector rather than Cartesian format (Downs and Priestnall 1999). Additionally, bank undercutting cannot be included in a conventional DEM, as that would require the allocation of more than one elevation to one location (Yao *et al.* 2011). Conversely, remote sensing can contribute to fluvial geomorphology in cases where the accuracy and precision of the sensed data meet the requirements for a specific investigation (Legleiter and Roberts 2009). For instance, drainage ditches, old river bends and channel dynamics may now be detected with moderately accurate LiDAR data (Notebaert *et al.* 2009). Some assumptions may be necessary to categorize each parcel (Yao *et al.* 2011). For instance, dense vegetation cover can prevent the detection of the underlying terrain (Resop and Hession 2010; De Rose and Basher 2011), which can necessitate data treatment. Additionally, the detection of submerged surfaces still represents a challenge, although passive optical remote sensing can be employed to predict bathymetry in shallow, clear-water rivers (Legleiter *et al.* 2009) and procedures exist to assess the accuracy and precision of bathymetry predictions (Legleiter and Roberts 2009).

Despite these limitations, the availability of remotely sensed data and the development of spatial analysis techniques, combined with the greater accessibility to computers of increasing power, have provided new ways to deal with spatial data. The integration of modelling with GIS and remote sensing is therefore very useful and promising. In general, it is recommended that modellers gain the relevant skills for using GIS and remote sensing software.

17.7 Physical models

A full review of physical models or hardware models is beyond the scope of this chapter. However, they are undoubtedly valuable modelling tools and, for completeness, we provide a brief overview of their characteristics. More detailed reviews of physical models are provided by Ashmore (1982), Schumm *et al.* (1987), Shen (1991), Ashworth *et al.* (1994), Peakall *et al.* (1996) and Warburton and Davies (1998), among others (see also Chapter 20).

Physical models in fluvial geomorphology are scaled hardware representations of an external fluvial setting – usually set up in a laboratory in the form of a flume, a sediment tank, a sand bed or a rain simulator where different experiments can be run. Physical models have been used to gain insight in a wide range of problems in fluvial geomorphology, ranging from simulation of small-scale flow and sediment transport dynamics (Lajeunesse *et al.* 2010; Frey and Church 2011) and channel bed evolution (Mao *et al.* 2011), to reach scale channel dynamics in meandering (Friedkin 1945; Braudrick *et al.* 2009; Visconti *et al.* 2010) and braided rivers (Ashmore 1982; Ashworth *et al.* 1994; Gardner and Ashmore 2011), to large-scale depositional structures such as alluvial fans (Clarke *et al.* 2010) and deltas (Sheets *et al.* 2002; Martin *et al.* 2009), to catchment-scale drainage evolution (Schumm *et al.* 1987; Hasbargen and Paola 2000; Bonnet and Crave 2003; Ouchi 2011).

The main strength of physical models is that, much like numerical or analytical models, they offer the advantage of experimental control. Thus, by systematically altering controlling variables, the response of a system to changes in single variables can be studied. An additional advantage is that, unlike the mathematical models, there is no concern about the suitability of the governing equations or algorithms, since the physical laws and processes that govern flow and sediment transport in the real world are by default also physically present in the model. However, non-linear scaling effects within these physical processes prohibit a direct translation between scales, such that care must be taken that the scaled hydraulics and sediment transport in the model correctly reflect the real-world physics (Peakall *et al.* 1996; Postma *et al.* 2008; Cooper and Tait 2009). This scaling problem becomes an even greater concern when vegetation or animal effects are considered (Gran and Paola 2001; Tal *et al.* 2004; Rice *et al.* 2009).

One particular point worth mentioning is the possible interplay between physical models and mathematical models. Physical models can provide a controlled dataset that can be used as a reference for the calibration and validation of numerical models, because the laboratory setting typically allows for all required data (i.e. initial conditions, forcing conditions and final result) to be measured in detail. Indeed, many numerical models are calibrated or validated using this approach (e.g. Van de Wiel and Darby 2004; Michaelides and Wainwright 2008). Conversely, predictions or inferences derived from numerical models, particularly the more surprising ones, could be tested *a posteriori* in a physical model. Such testing is currently not often done, possibly for financial considerations, but would potentially be one of the stronger types of validation of numerical models. Thus, mathematical models and physical models should not be seen as competing alternatives, but rather as complementary approaches to understanding fluvial geomorphology.

17.8 Overview of the modelling process

The use of any category of model involves a number of basic steps, the most important ones being data collection or acquisition, model calibration and model validation. These steps, which are described in this section, aim to ensure that a model sufficiently represents physical reality before it can be employed to understand the modelled system or make predictions regarding it.

All models require input data. In the case of fluvial models, these usually include topography, discharge and bed roughness.

Additional input requirements are dependent on the application and can involve elements such as bank and floodplain roughness, vegetation and infiltration rate. It seems trivial to note that the accuracy of the input data influences model output. Nonetheless, this is a point of importance as some data may be difficult or expensive to obtain accurately. Moreover, nearly all field data that are used as input are obtained from a relatively sparse collection of point measurements. Spreading these values over the spatial grid requires assumptions about their spatial (and sometimes temporal) distribution and usually involves some sort of interpolation routine, which introduces yet another source of uncertainty (see Table 17.3).

Most models require parameters to be set. A parameter is a variable which can be altered between model simulations, but which is kept constant for any given simulation. Parameters can be physical properties that are difficult to measure accurately (e.g. friction or soil moisture) or can be purely numerical components (e.g. convergence tolerance or smoothing kernel size). Often the values for these parameters can only be guessed at within a certain range and within that range any value is acceptable. When comparing the model results with a known data set, these parameters can be adjusted freely until an acceptable agreement between model prediction and observed data is found. This process is known as model calibration. The parameter values used to obtain the optimal result are then usually recommended to be used in other simulations with the same model. This calibration process is not undisputed (Beven and Binley 1992; Bates *et al.* 1998). Particularly when several parameters are adjusted during calibration, the uniqueness of an optimal setting is not guaranteed. There might be other combinations of calibration parameters that result in equally acceptable predictions, a condition known as 'model equifinality' (Beven 1996). Furthermore, the transferabilty of the calibrated parameters to other simulation scenarios (e.g. other spatial locations) often is questionable. Finally, the calibrated parameters obtained may mask systematic errors in model predictions or model structure. Alternative calibration schemes that partially address these problems have been proposed (Beven and Binley 1992; Bates *et al.* 1998; Hankin and Beven 1998; Campbell and Cox 1999). It should be noted, however, that the influence of the calibration parameters on the model results can be outweighed by other sources of uncertainty, such as grid construction (Hardy *et al.* 1999), or the limited accuracy of input data.

Once calibrated, a model is validated by running it against another known data set and checking the predicted results versus observed data. If this comparison is satisfactory, the model is said to perform well; if not, the model will be checked for errors and recalibrated. Both calibration and validation require complete data sets, in which some entity, for which predictions can be made, is known. The existence of, or access to, such data sets for natural fluvial systems is not always guaranteed (Bates *et al.* 1997). Very often, therefore, models are calibrated and validated using laboratory data, which may undermine their applicability in natural systems (ASCE Task Committee 1998).

17.9 Modelling applications in fluvial geomorphology

Modelling tools are used in a wide range of applications in fluvial geomorphology. Examples of practical modelling applications include prediction of the impacts of flushing flows on aquatic habitat (Milhous 1998ab), prediction of erosion and sedimentation impacts on land loss (Lohnes 1991) and planform adjustment (Mosselman 1995), in addition to the prediction of scour in the vicinity of bridges and other river

Table 17.3 Inherent limitations of fluvial geomorphological models (Haff 1996). Adapted from Haff, 1996.

Limitation	Examples
Model imperfection	Model imperfection refers to the fact that incremental 'improvement' in models at the laboratory scale do not necessarily add to our ability to make predictions at larger scales. For example, sediment transport is difficult to predict because the uniqueness of each natural sediment bed makes model implementation increasingly difficult as the model becomes more 'realistic' (e.g. Gomez and Church 1989)
Omission of significant processes	The larger the scale of the fluvial system, the greater is the chance that more than one important process will be omitted. For example, Tetzlaff and Harbaugh (1989) used a fluvial sedimentation model to simulate alluvial fan evolution, but in some locations debris flows may dominate fan construction (e.g. Whipple and Dunne 1992)
Unknown initial conditions	Initial conditions refer to the distribution of grain sizes, bank material characteristics, bed topography, etc. These conditions are often known only approximately or in some cases not at all. As far as predictive power is concerned, local site-specific data collection can be at least as important as model choice or model refinement (e.g. ASCE Task Committee 1998)
Sensitivity to initial conditions	Fluvial systems are non-linear, so there can exist a sensitivity to initial conditions that effectively prohibits detailed prediction of system evolution (e.g. Howard 1994; Howard *et al.* 1994; Murray and Paola 1994). A second issue is that of equifinality. In complex models, distinct parameters sets can lead to equally acceptable outcomes (Beven and Freer 2001)
Unresolved heterogeneity	In large-scale fluvial systems it may be impossible to define a meaningful averaging volume for each computational cell. Heterogeneity appears through in factors such as vegetative cover, soil type, etc.
External forcing	In fluvial systems, external forcing may be due to increases of discharge resulting from storms or dam releases. Predictive capabilities are limited if unpredictable external forcing can occur. In many cases, forcing can only be incorporated statistically if the distribution of events is known (e.g. Patton and Baker 1977)

Table 17.4 Generic indicators of model quality.

Fundamental questions	Criteria	Comments
What should a model provide?	Understanding	Models help to promote deeper understanding of the natural environment and underpin dialogue between the development of theory and critical experiments
	Forecasting potential	Forecasting is important in practical applications and as a means of testing the validity and range of understanding
What makes a good model?	Physical basis	When models have a strong physical basis, this provides consistency with other theories. This supports their validity and provides more users within the scientific community
	Simplicity	Models should be as simple as possible, so that they can be understood and communicated. It is difficult to construct a model in which more than three dominant processes interact at a time
	Generality and richness	Good models should be transferable to other geographical areas. Richness refers to the net information gain of the model
	Potential for scaling up or down	There is usually scope for model application over a range of scales only if the model has an explicit physical basis
What makes a good modelling tool?	Documentation	A well-documented user guide includes sufficiently detailed information on the functionalities that are available and on how to operate the modelling tool. It may also include sample cases, theoretical information on how the simulated processes are handled, the assumptions taken, etc. In cases where a model user can add self-programmed functionalities, a reference manual related to the program structure and libraries should be provided
	Support	A user may seek support if they wish to understand certain functionalities (even after reading the user guide) or in the case of programming errors. Support can be provided by the manufacturer or by a community of users. In the case of open-source tools, the user may view the programming code and modify it if required. The frequency at which the software is updated should also be considered
	Portability	A software that can be installed on computers with different architectures provides the user with greater flexibility in terms of work environment and desired performance

Adapted from Kirkby, 1996.

crossing infrastructure (Melville and Sutherland 1988; Johnson *et al.* 1999). The existence of a wide range of modelling tools (Section 17.2) implies that the selection of an appropriate model, geared to the demands of a specific application, is a difficult task. However, there are certain generic indicators of model quality, and also inherent limitations, that can be used as reference points in selecting from this diverse range of modelling tools.

Kirkby (1996) argues that a good model is characterized by an explicit physical basis, simplicity, generality, richness and the potential for scaling up or down. Table 17.4 summarizes these criteria, providing a framework that may be helpful when attempting to assess whether or not a particular model reaches acceptable minimum standards. However, the criteria should not be used to determine whether an individual model is 'better' or 'worse' than another model developed by a different author. Such comparative assessment exercises are very difficult to undertake, especially for complex numerical simulation models, and, unless considerable care is taken, the results of such exercises can be rather arbitrary and misleading. For example, the ASCE Task Committee (1998) tested discrete numerical models of bank erosion and channel widening developed by several different authors (Pizzuto 1990; Wiele 1992; Li and Wang 1993; Kovacs and Parker 1994) using a common data set obtained from a laboratory study (Ikeda 1981). The Task Committee deliberately avoided assessing the relative performance of the various models because they deviated in the numerical values of empirical coefficients used in the various process equations used by each author.

Table 17.4 indicates that model quality is the product of a series of criteria and is not simply dependent upon the formulation of complete, appropriate and realistic process laws, important though this step is. In fact, models consist of process laws or predictive rules, and also a set of input data to characterize the river reach or system for a specific application. Model quality is, therefore, highly constrained by the quality of the input data, in terms of both accuracy of data acquisition and the spatial scale at which the data are, or can be, acquired. A good example of the latter is provided by developments in the use of CFD modelling to investigate fluid flow processes in river channels (e.g. Bates *et al.* 1998; Lane 1998; Lane and Richards 1998) (see Section 17.5). In the development and application of CFD models, there has been a trend to increase the spatial resolution (the number of cells representing the spatial area of interest) in the expectation of improved insights into temporal and spatial processes (Hardy *et al.* 1999). Unfortunately, the spatial resolution at which a CFD model is applied affects the solution of the equations and thus the simulation results. Furthermore, data are rarely available at a sufficient level of detail to provide a data value for each cell in a numerical model, so that many values must be interpolated, leading to errors (Kirkby 1990).

Issues of data quality and quantity are implicit within several of the criteria listed in Table 17.4. For models with a low physical basis, parameter values are usually calibrated by model optimization, whereas models with a stronger physical basis may utilize parameters that are universal across a wide range of theory (Kirkby 1996). Models with a weak physical basis therefore require large data sets in order to calibrate parameter values and achieve a level of accuracy acceptable for forecasting purposes. Complex models generally have high demands on input data, but large numbers of parameters tend to provide opportunities for achieving a good fit between simulated and observed data. This includes cases where it is qualitatively plain that the 'right' answer is being produced by the 'wrong' set of processes (Kirkby 1996). Selecting a model with an optimal 'quality' level often involves achieving a balance between providing a strong physical basis and reducing the model complexity, in order to optimize input data quality. In trading off the increased physical basis of a model with its increasing complexity, a useful notion is the concept of model 'richness'. Model richness is analogous to the net information gain of the model (Kirkby 1996). For example, some complex and highly distributed landscape evolution models (e.g. Tucker and Slingerland 1997) have very large input data requirements, but when used to forecast the sediment yield at the outlet of the watershed their net information gain is strongly negative.

For many practical applications, particularly those where the level of expertise of the user is low relative to the complexity of the model, practical criteria additional to those listed in Table 17.4 may also be significant factors in assessing the overall 'quality' of a model. In particular, model portability, accessibility, cost and usability are, in practice, often just as important as the criteria listed in Table 17.4. These issues are almost exclusively concerned with quantitative models, which are usually implemented and used in the form of computer software packages. It is incumbent upon the developers of such packages to ensure that the modelling software is readily available (e.g. via the Internet), well documented and portable across a range of different hardware platforms. This is important in facilitating scientific dialogue and exchange and in promoting the dissemination of modelling tools for practical applications. Furthermore, model developers and users need to consider whether or not data entry and interpretation of output data are facilitated by a friendly graphical interface.

In determining minimum standards of model quality, it is important to recognize that there are inherent limitations to fluvial modelling. Several authors have questioned the nature of prediction in the geosciences in general (Tetzlaff 1989; Oreskes *et al.* 1994; Cleland 2001) and in fluvial geomorphology (Baker 1988, 1994) and hydrology (Anderson and Woessner 1992; Konikow and Bredehoeft 1992; Rojstaczer 1994), in particular. There are several distinctive features of geomorphological systems that make prediction inherently difficult (Haff 1996). Table 17.3 summarizes these sources of uncertainty and error as they apply to quantitative models in fluvial geomorphology.

Beven (1996) also discussed problems of equifinality and uncertainty in geomorphological modelling. The problem of equifinality in this context refers to modelling scenarios wherein agreement between modelled and observed data can be obtained by a wide variety of parameter sets. As a result, this leads to uncertainty in inference and prediction (Beven 1996; Beven and Freer 2001). It will be apparent that this problem is closely related to the issue of model complexity described above, wherein equifinality is associated primarily with complex models that have complex input data requirements. To overcome this difficulty, carefully designed validation and verification exercises are required to demonstrate that agreement between observed and modelled data is assessed both for (non-distributed) bulk parameters and for spatially distributed parameters (see Section 17.5).

Although the above discussion perhaps provides some guidance on acceptable minimum standards for modelling tools, potential users are still faced with the problem of selecting a modelling tool appropriate for a specific application. The wide range of models associated with the treatment of practical fluvial geomorphological problems makes it essential that practitioners adopt a broad and rational approach to such problems (ASCE Task Committee 1998). Such an approach is needed to analyse the majority of problems that arise with the assurance that important factors are not overlooked, appropriate techniques are applied and hence effective solutions are developed.

17.10 Generic framework for fluvial geomorphological modelling applications

The generic framework presented here is based on a procedure recommended by the ASCE Task Committee (1998). Although this Task Committee was concerned with the specific context of modelling river width adjustment, it is still a valid approach that is based on amassing and utilizing a range of methods and techniques appropriate to a specific problem. This framework recognizes that although each case is unique, there are a number of generic elements that are relevant for the majority of modelling situations (ASCE Task Committee 1998). Whereas the methodological steps of the modelling procedure are usually formulated for the application of a single model [for an example, see the steps proposed by Refsgaard (2000) for hydrological models], the generic framework allows for sequentially using models of different categories in order to examine a single problem (Fig. 17.1). Also, the generic framework assumes that the selected model(s) is (are) suitable for the investigated phenomenon and hence that no code programming/verification is necessary.

The first of these steps, problem identification, is perhaps the most crucial in that it involves the formulation of a clear set of objectives for the modelling application. A fundamental part of problem identification involves determining the resources (time, money, skill level of personnel, etc.) available for the project. The problem should be formulated in terms of whether the modelling objectives are geared towards understanding existing behaviour or whether predictions of future system behaviour are required. For practical applications, key questions relate to the definition of who or what are affected by the 'problem' and what level of analysis and response is appropriate.

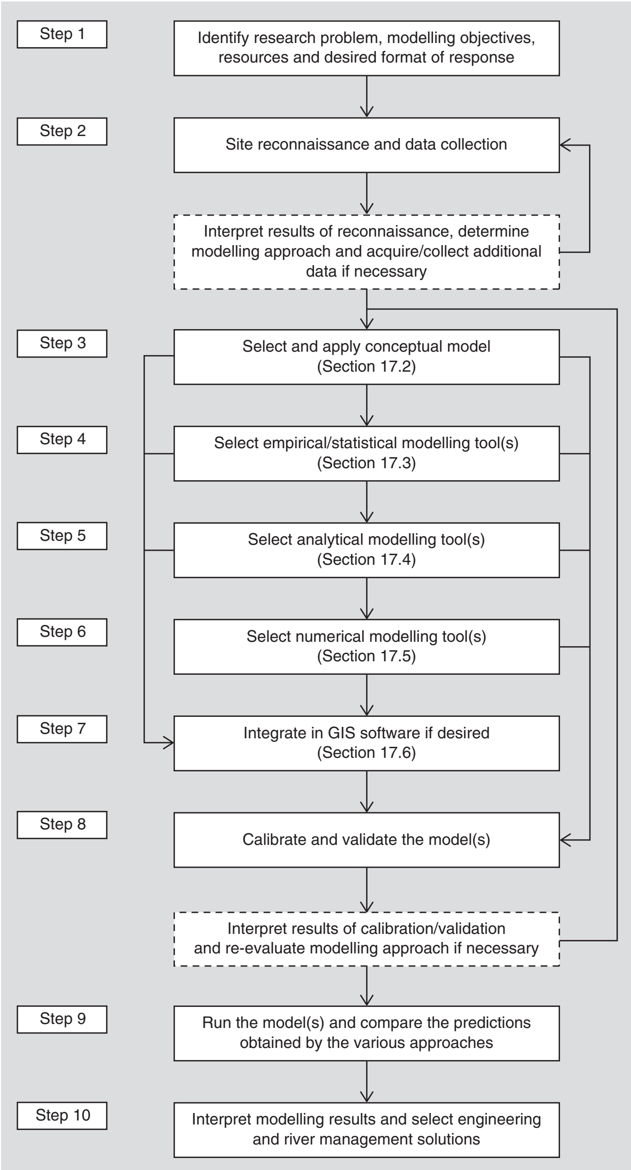


Figure 17.1 Proposed generic framework for applying modelling tools in practical applications.

The second step (Fig. 17.1) may, if appropriate, comprise a reconnaissance visit to the site and river reaches located upstream or downstream (ASCE Task Committee 1998). This step allows the identification of channel characteristics, bank conditions, bank materials, extent of existing or expected bank erosion problems, nature of the flow and bed materials, presence and nature of any vegetation and presence and condition of any engineering structure. Appropriate stream reconnaissance techniques have been described by, among others, Kellerhals *et al.* (1976), Downs and Brookes (1994) and Thorne *et al.* (1996). Reconnaissance should be viewed as the necessary means by which the modeller familiarizes themself with the characteristics of the problem, as they are manifest in the field. Reconnaissance can also help to determine problem complexity and the aspects of the problem that might necessitate the selection of specific modelling tools in subsequent steps of the framework. Reconnaissance is, therefore, a key step in developing a strategic view of the problem and planning an appropriate modelling strategy. The data required for the preliminary site assessment, and subsequently for parameterizing, calibration, validating and using the model(s), can be acquired or collected (if these are not readily available) if financial resources permit. Restrictions on the availability, type and resolution of data can determine the category of model that can correctly be employed to investigate or predict phenomena at a certain site of interest.

After reconnaissance, the user addresses the problem through the application of modelling tools progressively increasing in complexity (Fig. 17.1). This allows the modeller to absorb manageable increments of knowledge at each step of the process. This step-by-step approach is flexible in its application, depending on the complexity of a particular problem. For very simple problems, users may decide that simple models (Steps 3 and 4) are sufficient to solve the problem without the need for unnecessary investment in complex modelling approaches (Steps 5 and 6). However, for complex problems every step would be used, as understanding developed from application of complex simulation models is often enhanced by attempts to break down the system to its simplest possible components.

Step 3 therefore involves the selection of an appropriate conceptual model (Section 17.2). Conceptual models aid in the identification of dominant processes and trends and help in forming a structure for subsequent, more detailed, modelling. Following on from this, Step 4 consists of the selection and application of relatively simple empirical or statistical models (Section 17.3). Step 4 differs from previous steps in that quantitative, rather than qualitative, predictions are obtained. If the complexity and severity of the problem merit further analysis, Steps 5 and 6 involve the selection of analytical (Section 17.4) and numerical simulation (Section 17.5) models, respectively. Step 7 involves integration, where desired, of the selected model in a GIS framework (Section 17.6).

The application of quantitative modelling tools in Steps 4–6 normally requires the model(s) to be calibrated and validated (Step 8; Section 17.8) against available field data. Assuming that the calibration and validation procedures are achieved successfully, the model is said to be reliable for a particular context and is ready to be made in predictive mode. Users can then run the model(s) to assess the outcome of defined scenarios, compare the predictions obtained from various approaches (Step 9) and interpret the results (Step 10). If the user is undertaking a practical modelling application, appropriate engineering solutions or management strategies can then be selected and implemented while considering the knowledge gained about the investigated river system during the modelling process.

17.11 Case study: meander dynamics

The operation of the generic framework (Fig. 17.1) can be illustrated using an example of a typical 'problem', that of meander migration dynamics.

Although all rivers are worthy of investigation, meandering rivers have been the subject of extensive research, due to their common occurrence, aesthetic appeal and the intriguing question as to the mechanism of formation of the sinuous planform. Furthermore, as meandering rivers migrate across their floodplains, erosion of the outer banks can result in loss of agricultural land, highways, buildings and other infrastructure. In terms of problem identification (Step 1 in Fig. 17.1), there is a clear priority in furthering our understanding of the process–form linkages between the fluid flows driving morphological change and the resulting meander migration rates. Particularly important in this regard is the relationship between meander morphology, flow hydraulics and meander migration rate. This example problem is based, therefore, on the pure scientific application of modelling tools. In this context stream reconnaissance (Step 2 in Fig. 17.1) is not a necessary component of the proposed generic framework and we may pass directly on to application of relevant conceptual models (Step 3 in Fig. 17.1).

The nature of channel migration is well represented in conceptual models. Empirically based conceptual models describe sequences and styles of change, based on observations from aerial photographs or historical maps (see Chapter 4). Hooke (1997) provided a detailed overview and distinguished four main styles of change in meandering channels: migration, confined migration, growth and compound development and cut-off. These empirically based, conceptual models provide a clear visual interpretation of the evolution of meanders, but provide little process explanation. The first physically based conceptual models were derived by Thomson (1876) and Einstein (1926), who recognized the role of helical flows in meander bends. Current physically based conceptual models provide some further detail about the processes involved (Thompson 1986): migration is a consequence of erosion of banks through fluvial entrainment of bank sediments, possibly leading to undercutting and mass failure of the upper bank. The fluvial entrainment is enhanced at the outer bank as secondary flows in meander bends redistribute flow velocities across the channel with faster flow velocities near the outer bend.

Empirical models (Step 4 in Fig. 17.1) have been used to explain meander migration in terms of large-scale flow separation processes in bends (Markham and Thorne 1992). Flow separation occurs where boundaries turn away from the main flow, causing the streamlines to diverge. The exact point at which the flow separates is closely linked to the shape and roughness of the boundary (Markham and Thorne 1992). In particular, a strong relationship between the onset of flow separation in pipes and the ratio of radius of curvature to width (R/W) has been found (Bagnold 1960). As R/W declined to about 2, inner bank separation was found to occur, altering the apparent geometry in such a way that the flow resistance decreases to a minimum value. In rivers, separation occurs at the inside bank downstream of the apex of the point bar and on the outside of the channel close to, but upstream of, the bend apex (Carey 1969; Leeder and Bridges 1975). Hickin and Nanson (1975) and Hickin (1977, 1978) used these observations to develop an empirical model of meander migration that relates migration rates to the value of R/W (Fig. 17.2). Their data from the Squamish and Beatton rivers in British Columbia show that meander migration rates have a maximum value associated with R/W values around 2–3. This is consistent with the observed correspondence of minimum friction factor related to flow separation. This empirical model therefore appears to have a reasonably solid theoretical basis and has the potential to be a useful tool when calibrated for a particular stream system. Since the mid-1980s, however, emphasis has shifted to first analytical and later numerical models of meander dynamics.

Analytical models (Step 5 in Fig. 17.1) are based on a linearization of the flow equations. Several such models have been developed (Ikeda *et al.* 1981; Smith and McLean 1984; Blondeaux and Seminara 1985; Johannesson and Parker 1989; Odgaard 1989; Zolezzi and Seminara 2001), which differ mainly in the treatment of secondary flows and in the interaction with bed topography. The main common feature is the treatment of the lateral meander migration, which is calculated as

$$\xi = \varepsilon(U + U_b) \tag{17.4}$$

where ξ is the outer bank erosion rate, U is the mean flow velocity, U_b is the near-bank flow velocity and ε is a proportionality coefficient whose value is determined by calibration. This relation was proposed independently by Hasegawa (1977; cited in Parker *et al.* 2011) and Ikeda *et al.* (1981) and is now referred to as the HIPS equation (Parker *et al.* 2011). It is not undisputed, as it simplifies the migration process by combining fluvial and geotechnical bank erosion in a single proportionality coefficient, ε, which is considered to be a 'catch-all' parameter representing the relative erodibility of the bank materials. Its value must be determined by calibration, which requires

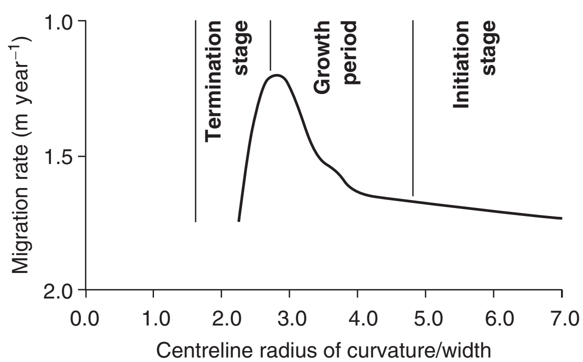


Figure 17.2 Meander migration rate and evolutionary stage as a function of radius of curvature to width ratio. Source: Markham and Thorne, 1992. Reproduced with permission of Wiley.

detailed historical data on meander planform changes, but these are often not readily available. Moreover, most models assume that the erodibility coefficient is constant both through space and time. This is most unlikely in the context of natural meandering streams, which create erodible point bars and resistant fine-grained deposits (in oxbow lakes) as they migrate across the floodplain. In practice the proportionality coefficient, ε, is therefore often seen as a 'fudge factor' in the calibration of the models (Parker *et al.* 2011). Additionally, the models require several assumptions be made in the linearization of the flow equations. For example, the models assume constant discharge, constant channel width, mild curvature and mild curvature changes (Seminara 2006; Camporeale *et al.* 2007; Blanckaert and De Vriend 2010). Two essential limitations that follow from these assumptions are, first, that the models represent only a limited range of meandering rivers, and, second, that only relatively short-term evolution (up to meander cut-off) can be studied analytically. However, despite these limitations, the main advantage of these linear models is that they offer a tractable solution to the meander evolution problem. Application of these linear models has led to a number of key insights. They have shown how straight channels evolve into meandering channels given a small initial perturbations in the straight channel (e.g. a small asymmetry in the channel bed at a cross-section or a slight cross-sectional asymmetry in the incoming sediment load at the upstream boundary), how both symmetrical and skewed meanders can form (Johannesson and Parker 1989; Chen and Duan 2006), how alternate bars form in and interact with meandering channels (Blondeaux and Seminara 1985; Colombini *et al.* 1987; Seminara and Tubino 1989) and how mid-channel bars form and interact with meandering channels (Luchi *et al.* 2010). However, the arguably most important insight is that these models have shown that meanders behave like oscillators, which can resonate with the channel bed. Resonant meanders have steady alternate bars and do not migrate, sub-resonant meanders migrate downstream and super-resonant meanders migrate upstream (Blondeaux and Seminara 1985; Seminara *et al.* 2001; Camporeale and Ridolfi 2006; Seminara 2006).

Numerical models of river meandering (Step 6 in Fig. 17.1) fall broadly in one of four categories. The first of these is a direct numerical implementation of the linear analytical models (e.g. Howard and Knutson 1984; Johannesson and Parker 1989; Howard 1992, 1996; Meakin *et al.* 1996; Sun *et al.* 1996, 2001; Stølum 1998; Zolezzi and Seminara 2001; Frascati and Lanzoni 2009; Xu *et al.* 2011). The channel is divided into a number of segments. Given the curvature and topography of the channel, the mean flow velocity, U and the near-bank flow velocity, U_b, are calculated. Each segment is then moved laterally according to the resulting lateral migration (eqn. 17.4), giving rise to a new centreline and a new curvature. This calculation sequence is then repeated for the next time step. This numerical approach has the advantage that the meander evolution can be simulated and visualized over longer time frames, e.g. decades or centuries, which cannot be done with the analytical models. However, they do suffer from many of the analytical models' limitations: simplified representation of secondary flows, assumed constant width, mild curvature, mild curvature gradients, no tributaries, reliance on a homogeneous sediment structure. Also, notably, they do rely on eqn. 17.4 to calculate lateral migration and therefore suffer from the same limitations. Some efforts have been made to overcome these deficiencies partially, for example, by allowing the local erodibility, ε, to vary (Sun *et al.* 1996; Posner and Duan 2012), by accounting for heterogeneous sediment (Sun *et al.* 2001), by allowing the inner bank to migrate at different rates to the outer bank and thus fully bypassing the constant width assumption (Parker *et al.* 2011), or by accounting for bank height in the lateral erosion process (Xu *et al.* 2011).

The second category of numerical models is similar in structure to the previous type, but relies on non-linear approximations of the flow equations (Zolezzi and Seminara 2001; Lancaster and Bras 2002; Camporeale *et al.* 2007; Bolla Pittaluga *et al.* 2009; Blanckaert and De Vriend 2010). These non-linear models, which cannot be solved analytically, have an improved representation of secondary flows, resulting in a more realistic flow model and allowing the curvature assumption to be relaxed. The models can therefore be used in a wider range of scenarios than their linear counterparts. However, they still assume homogeneous sediment structure and constant width [but see Luchi *et al.* (2011) for a variation which allows for oscillating width adjustments] and they still rely on eqn. 17.4 to calculate lateral migration and thus inherit all the problems associated with that equation. Nonetheless, these models have furthered the understanding of meander dynamics by showing how compound meander bends form (Seminara *et al.* 2001; Lancaster and Bras 2002), how meanders of small wavelength decay whereas meanders of larger wavelength grow to cut-off stage (Edwards and Smith 2001, 2002), how the how cut-offs affect the evolution of meander belts (Camporeale *et al.* 2005, 2008; Frascati and Lanzoni 2009) and how vegetation affects the meander dynamics (Perucca *et al.* 2007). Additionally, both the linear and non-linear numerical models have been used to investigate the non-linear dynamics of long-term meander evolution (Stølum 1996, 1998; Hooke 2007; Frascati and Lanzoni 2010; Bolla-Pittaluga and Seminara 2011; Xu *et al.* 2011) (Fig. 17.3).

The third category of numerical models takes a more reductionist approach, where flow is modelled through a fully non-linear 2D or 3D computational fluid dynamics algorithm and where sediment transport and bank erosion are explicitly represented using physically based submodels. Most CFD models for simulating meander bend flow focus on calculating the flow field only (Hodskinson and Ferguson 1998; Lien *et al.* 1999; Blanckaert and De Vriend 2003) or use the calculated flow field to compute sediment fluxes and bed topography change (Rüther and Olsen 2005; Wu *et al.* 2005; Fischer-Antze *et al.* 2009; Mekonnen *et al.* 2010; Vasquez *et al.* 2011). Further calculations of bank erosion and channel migration are more challenging, since they typically require updating the grid over which the

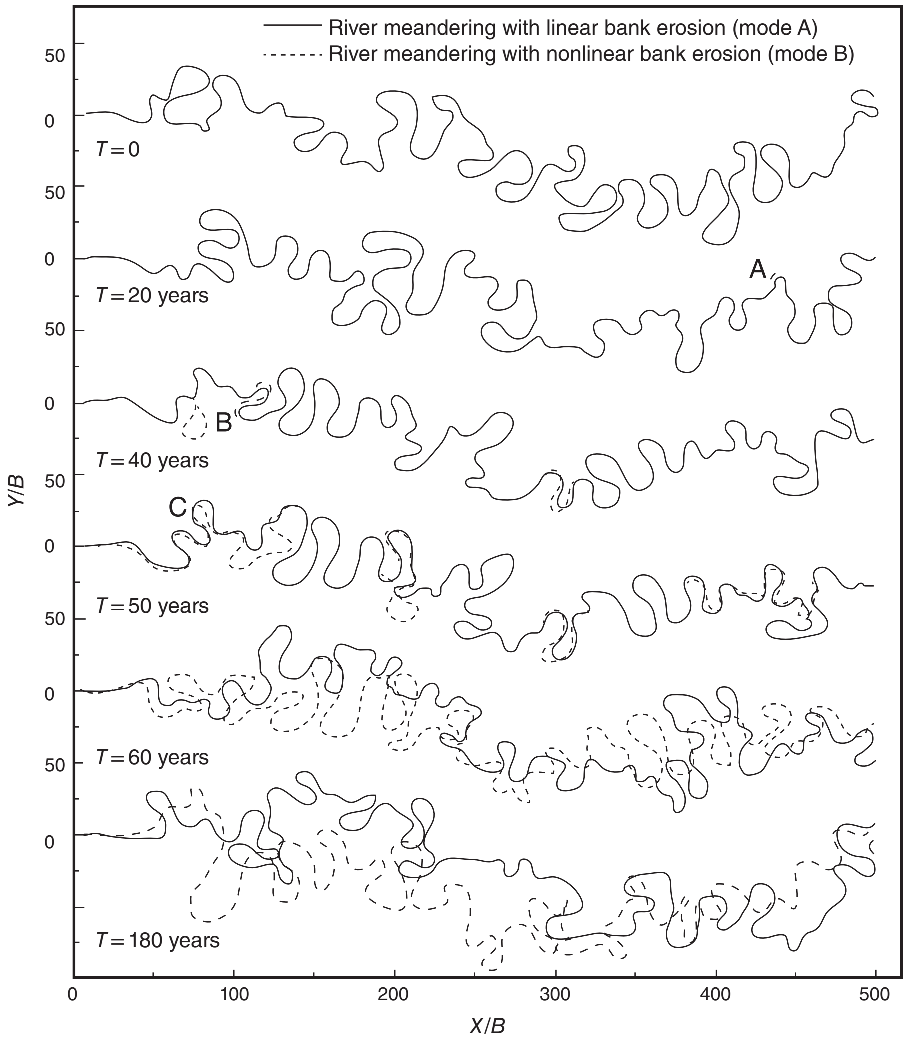


Figure 17.3 Comparison of meander evolution in a linear and non-linear meander model. Source: Xu *et al*, 2011. Reproduced with permission of Elsevier.

calculations are performed. However, some advances have been made in this regard (Mosselman 1998; Nagata *et al.* 2000; Darby *et al.* 2002; Olsen 2003; Duan and Julien 2005, 2010). These compound models typically utilize a three-step solution procedure. In the first step, the flow is computed while keeping the bed and bank configuration fixed. Sediment transport fluxes and bank erosion rates resulting from the predicted flow field are then computed. Second, bed level changes are computed from the sediment transport flux gradients and the input of bank erosion products. Finally, bankline changes are computed from the bank erosion rates (Fig. 17.4). The predictive ability of these models is, therefore, related in part to the predictive abilities of each of the submodels used in these three modules. In fact, limitations with these models are to be expected in that they utilize empirically calibrated sediment transport equations and they commonly utilize a bank erosion submodel tailored for specific physical environments. For example, the model of Mosselman (1998) is tailored to cohesive banks, whereas that of Nagata *et al.* (2000) simulates non-cohesive bank materials. Darby *et al.* (2002) included a geotechnical analysis of planar bank failure to simulate bank collapse. Nonetheless, these approaches represent the state of the art in simulating meander migration and offer considerable potential. This is because they offer generic frameworks into which improvements in individual submodels (e.g. improved flow models or improved bank erosion algorithms) can be readily introduced. These reductionist models are useful over relatively short time and spatial scales, but they require a great deal of input data, computational power and technical expertise. So far this has restricted their use in more general morphological models used to account for changes in channel planform resulting from bank erosion and accretion processes.

In contrast to the reductionist CFD models, the final category of numerical models follows the reduced complexity modelling paradigm. These models attempt to simulate the channel–floodplain interactions of meandering rivers in a

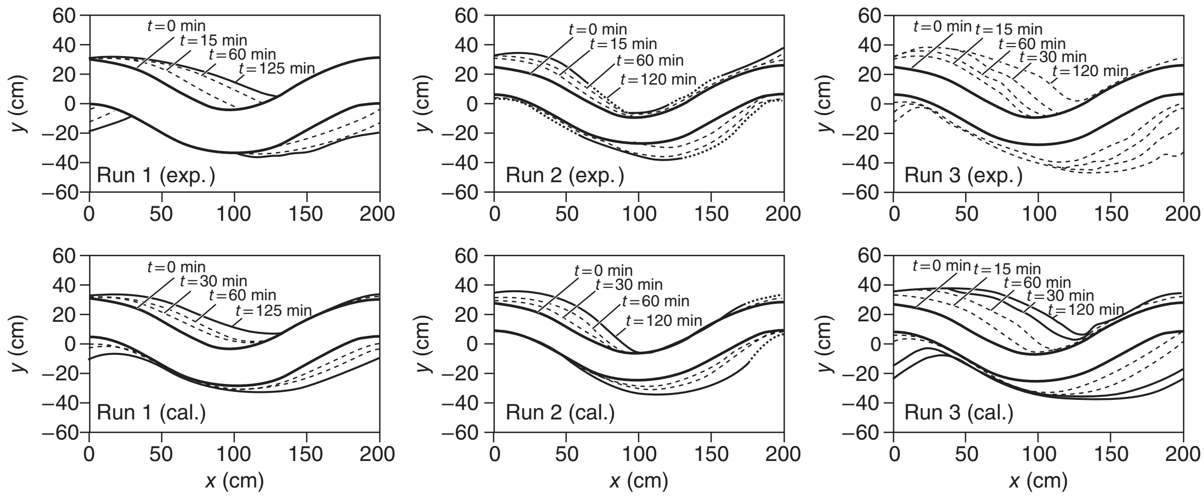


Figure 17.4 Comparison of observed channel migration for three laboratory flume experiments (upper plots) with numerical simulation results (lower plots). Source: Nagata *et al*, 2000. Reproduced with permission from ASCE Library.

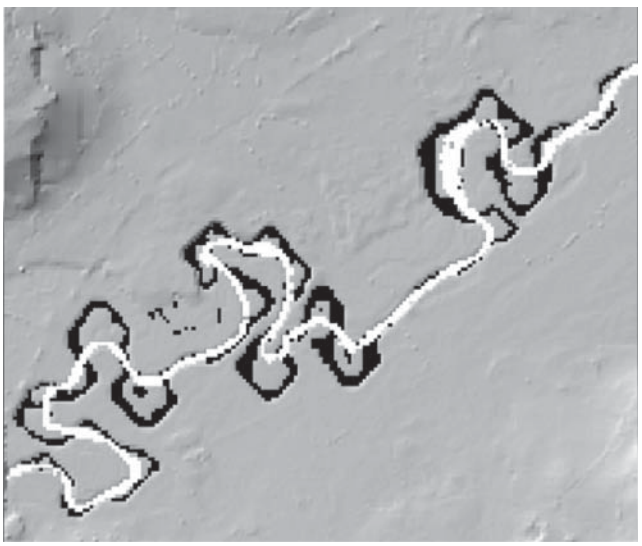


Figure 17.5 Meander migration simulated with a cellular automaton model. Initial channel position is in white, final channel position in black. From Coulthard and Van de Wiel (2006).

cellular automaton framework (Coulthard and Van de Wiel 2006, 2012) (Fig. 17.5). They are in a relatively early stage of development but results so far are encouraging (Coulthard and Van de Wiel 2012). The main advantage of this approach is that it can be readily integrated with existing cellular automaton landscape evolution models, thus offering the potential for simulating meander evolution in the context of catchment-scale environmental change, such as land-use change or climate change.

Few of these meander migration models have yet been rigorously tested using field data (Step 8 in Fig. 17.1). These models must, therefore, be regarded as being in an early stage of development and are, at present, modelling tools for research rather than engineering purposes (Step 10 in Fig. 17.1). For example, they have been used to investigate the impacts of vegetation on bank erosion and meandering (Van de Wiel and Darby 2004, 2007; Crosato and Saleh 2010). However, as hardware and software develop, it is likely that this type of model will, in the future, be properly calibrated and validated to be used more frequently for engineering-oriented morphological modelling problems.

17.12 Conclusion

This chapter has shown that modelling tools are used for a wide variety of both pure and practical applications in fluvial geomorphology. It is unsurprising, therefore, that many different types of models have been developed. These types of models were classified here into six main categories (conceptual or theoretical, statistical, analytical, numerical, GIS-based and physical models; Sections 17.2–17.7). Within each of these categories, a selection of models was reviewed, highlighting the strengths, weaknesses, capabilities and limitations of each discrete approach.

Because there is considerable diversity in the range of modelling tools, both within and between different categories, users are faced with significant practical difficulties when selecting a modelling tool or tools to address a specific application. Fortunately, certain generic indicators of model quality (Table 17.4) can be identified and these can be used to help decide if a particular model meets acceptable quality standards for a specific application. Similarly, certain inherent limitations of models can be identified (Table 17.3). A particular constraint on the overall quality of a specific model appears to be the amount of input data required to obtain a prediction, relative to the amount of output data obtained. Furthermore, the scale at which input data is or can be acquired, relative to the scale at which the model is applied, is another key limiting factor. These constraints mean that there is often (although not always) a trade-off between choosing models with a strong physical basis and complex data requirements and choosing models that have less demanding data requirements and a commensurately lower theoretical basis.

It is important to recognize that ranking the quality of individual models is difficult, time consuming and often ultimately flawed. Instead of providing recommendations regarding the selection of different types of fluvial geomorphological models, a flexible generic framework for model applications has instead been proposed. In this framework, clear problem formulation provides the basis for a rational modelling approach. For modelling applications involving relatively simple problems, conceptual or statistical models alone may be sufficient to generate reliable predictions and develop the required level of understanding of the problem. For more complex problems, application of a spectrum of modelling strategies ranging from simple to complex is required. The application of multiple modelling tools provides the user with overlapping sets of predictions that lead to an enhanced level of understanding, in addition to increased confidence in the predictions themselves.

Acknowledgements

M.J.VdW. is supported by an NSERC Discovery Grant. Y.Y.R. is supported through FQRNT and OGS scholarships.

References

Abernethy, B. and Rutherfurd, I.D. 2000. The effect of riparian tree roots on the mass-stability of riverbanks. *Earth Surface Processes and Landforms* 25: 921–937.

Abernethy, B. and Rutherfurd, I.D. 2001. The distribution and strength of riparian tree roots in relation to riverbank reinforcement. *Hydrological Processes* 15: 63–79.

Abrahams, A.D., Li, G. and Atkinson, J.F. 1994. Step-pool streams: adjustment to maximum flow resistance. *Water Resources Research* 31: 2593–2602.

ASCE Task Committee. 1998. River width adjustment. II. Modelling. *Journal of Hydraulic Engineering* 124: 903–917.

Anderson, M.P. and Woessner, W.W. 1992. The role of the postaudit in model validation. *Advances in Water Resources* 15: 167–173.

Ashmore, P.E. 1982. Laboratory modelling of gravel braided stream morphology. *Earth Surface Processes and Landforms* 7: 201–225.

Ashworth, P.J., Best, J.L., Leddy, J.O. and Geehan, G.W. 1994. The physical modelling of braided rivers and deposition of fine-grained sediment. In: Kirkby, M.J., ed., *Process Models and Theoretical Geomorphology*, Chichester: John Wiley & Sons, pp. 115–139.

Bagnold, R.A. 1960. *Some Aspects of the Shape of River Meanders*, US Geological Survey Professional Paper 282E, Washington, DC: US Government Printing Office.

Baker, V.R. 1988. Geological fluvial geomorphology. *Bulletin of the Geological Society of America* 100: 1157–1167.

Baker, V.R. 1994. Geomorphological understanding of floods. *Geomorphology* 10: 139–156.

Bates, P.D., Anderson, M.G., Price, D.A., Hardy, R.J. and Smith, C.N. 1996. Analysis and development of hydraulic models for floodplain flows. In: Anderson, M.G., Walling, D.E. and Bates, P.D., eds., *Floodplain Processes*, Chichester: John Wiley & Sons, pp.215–254.

Bates, P.D., Horritt, M.S., Smith, C.N. and Mason, D. 1997. Integrating remote sensing observations of flood hydrology and hydraulic modelling. *Hydrological Processes* 11: 1777–1795.

Bates, P.D., Horritt, M. and Hervouet, J.M. 1998. Investigating two-dimensional, finite element predictions of floodplain inundation using fractal generated topography. *Hydrological Processes* 12: 1257–1277.

Beffa, C. and Connell, R.J. 2001 Two-dimensional flood plain flow. *I. Model description. Journal of Hydrologic Engineering* 6: 397–405.

Benson, M.A. 1965. Spurious correlation in hydraulics and hydrology. *Journal of the Hydraulics Division, American Society of Civil Engineers* 91: 35–43.

Bettess, R., White, W.R. and Reeve, C.E. 1988. On the width of regime channels. In: White, W.R., ed., *River Regime*, Chichester: John Wiley & Sons, pp. 149–161.

Beven, K. 1996. Equifinality and uncertainty in geomorphological modelling. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 289–313.

Beven, K. and Binley, A. 1992. The future of distributed models: model calibration and uncertainty prediction. *Hydrological Processes* 6: 279–298.

Beven, K. and Freer, J. 2001. Equifinality, data assimilation and uncertainty estimation in mechanistic modelling of complex environmental systems using the GLUE methodology. *Journal of Hydrology* 249: 11–29.

Bishop, A.W. 1955. The use of the slip circle in the stability analysis of slopes. *Géotechnique* 5(1):7–17.

Blanckaert, K. and De Vriend, H.J. 2003. Nonlinear modeling of mean flow redistribution in curved open channels. *Water Resources Research* 39: 1375.

Blanckaert, K. and De Vriend, H.J. 2010. Meander dynamics: a nonlinear model without curvature restrictions for flow in open-channel bends. *Journal of Geophysical Research* 115: F04011.

Blondeaux, P. and Seminara, G. 1985. A unified bar-bend theory of river meanders. *Journal of Fluid Mechanics* 112: 363–377.

Bolla Pittaluga, M. and Seminara, G. 2011. Nonlinearity and unsteadiness in river mandering: a review of progress in theory and modelling. *Earth Surface Processes and Landforms* 36: 20–38.

Bolla Pittaluga, M., Nobile, G. and Seminara, G. 2009. A nonlinear model for river meandering. *Water Resources Research* 45: W04432.

Bonnet, S. and Crave, A. 2003. Landscape response to climate change: insights from experimental modeling and implications for tectonic versus climatic uplift of topography. *Geology* 31: 123–126.

Booker, D.J. and Dunbar, M.J. 2004. Application of physical habitat simulation (PHABSIM) modelling to modified urban river channels. *River Research and Applications* 20: 167–183.

Booker, D.J., Sear, D.A. and Payne, A.J. 2001. Modelling three-dimensional flow structures and patterns of boundary shear stress in a natural pool-riffle sequence. *Earth Surface Processes and Landforms* 26: 553–576.

Bradbrook, K.F., Biron, P.M., Lane, S.N., Richards, K.S. and Roy, A.G. 1998. Investigation of controls on secondary circulation in a simple confluence geometry using a three-dimensional numerical model. *Hydrological Processes* 12: 1371–1396.

Bradbrook, K.F., Lane, S.N., Richards, K.S., Biron, P.M. and Roy, A.G. 2001. Role of bed discordance at asymmetrical river confluences. *Journal of Hydraulic Engineering* 127(5): 351–368.

Brakenridge, G.R., Carlos, H. and Anderson, E. 2002. River gaging reaches: a strategy for MODIS-based river monitoring. *Proceedings of the Society of Photo-Optical Instrumentation Engineers (SPIE)* 4886: 479–486.

Bras, R.L., Tucker, G.E. and Teles, V. 2003. Six myths about mathematical modeling in geomorphology. In: Wilcock, P.R. and Iverson, R.M., eds., *Prediction in Geomorphology*, Washington, DC: American Geophysical Union, pp. 63–79.

Brasington, J. and Richards, K. 2007. Reduced-complexity, physically-based geomorphological modelling for catchment and river management. *Geomorphology* 90: 171–177.

Braudrick, C.A., Dietrich, W.E., Leverich, G.T. and Sklar, L.S. 2009. Experimental evidence for the conditions necessary to sustain meandering in coarse-bedded rivers. *Proceedings of the National Academy of Sciences of the United States of America* 106: 16936–16941.

Brett, M.T. 2004. When is a correlation between non-independent variables 'spurious'? *Oikos* 105: 647–656.

Brookes, A. and Sear, D.A. 1996. Geomorphological principles for restoring channels. In: Brookes, A. and Shields, F.D. Jr,, eds., *River Channel Restoration*, Chichester: John Wiley & Sons, pp. 75–101.

Brookes, A. and Shields, F.D. Jr,, eds. 1996. *River Channel Restoration*. Chichester: John Wiley & Sons.

Caamaño, D., Goodwin, P. and Buffington, J.M. 2012. Flow structure through pool-riffle sequences and a conceptual model for their sustainability in gravel-bed rivers. *River Research and Applications* 28: 377–389.

Campbell, E.P. and Cox, D.R. 1999. A Bayesian approach to parameter estimation and pooling in nonlinear flood event models. *Water Resources Research* 35: 211–220.

Camporeale, C. and Ridolfi, L. 2006. Convective nature of the planimetric instability in meandering river dynamics. *Physical Review E* 73: 026311.

Camporeale, C., Perona, P., Porporato, A. and Ridolfi, L. 2005. On the long-term behavior of meandering rivers. *Water Resources Research* 41: W12403.

Camporeale, C., Perona, P., Porporato, A. and Ridolfi, L. 2007. Hierarchy of models for meandering rivers and related morphodynamic processes. *Reviews of Geophysics* 45: RG1001.

Camporeale, C., Perucca, E. and Ridolfi, L. 2008. Significance of cutoff in meandering river dynamics. *Journal of Geophysical Research* 113: F01001.

Carey, W.C. 1969. Formation of flood plain lands. *Journal of the Hydraulics Division, American Society of Civil Engineers* 95: 981–994.

Carson, M.A. and Griffiths, G.A. 1987. Bedload transport in gravel channels. *Journal of Hydrology (New Zealand)* 26: 1–151.

Casagli, N., Rinaldi, M., Gargini, A. and Curini, A. 1999. Pore water pressure and streambank stability: results from a monitoring site on the Sieve River, Italy. *Earth Surface Processes and Landforms* 24: 1095–1114.

Chang, H.H. 1980. Geometry of gravel streams. *Journal of the Hydraulics Division, American Society of Civil Engineers* 105: 1443–1456.

Chang, H.H. 1988. *Fluvial Processes in River Engineering*, New York: Wiley-Interscience.

Chen, D. and Duan, J.G. 2006. Modeling width adjustment in meandering channels. *Journal of Hydrology* 321: 59–76.

Chin, A. 1999. The morphologic structure of step-pools in mountain streams. *Geomorphology* 27: 191–204.

Clarke, L., Quine, T.A. and Nicholas, A. 2010. An experimental investigation of autogenic behaviour during alluvial fan evolution. *Geomorphology* 115: 278–285.

Cleland, C.E. 2001. Historical science, experimental science and the scientific method. *Geology* 29: 987–990.

Clifford, N.J., Acreman, M.C. and Booker, D.J. 2008. Hydrological and hydraulic aspects of river restoration uncertainty for ecological purposes. In: Darby, S.E. and Sear, D.A., eds., *River Restoration: Managing the Uncertainty in Restoring Physical Habitat*, Chichester: John Wiley & Sons, pp. 105–138.

Colombini, M., Seminara, G. and Tubino, M. 1987. Finite-amplitude alternate bars. *Journal of Fluid Mechanics* 181: 213–232.

Cooper, J.R. and Tait, S.J. 2009. Water-worked gravel beds in laboratory flumes – a natural analogue? *Earth Surface Processes and Landforms* 34: 384–397.

Coulthard, T.J. and Van de Wiel, M.J. 2006. A cellular model of river meandering. *Earth Surface Processes and Landforms* 31: 123–132.

Coulthard, T.J. and Van de Wiel, M.J. 2012. Modelling river history and evolution. *Philosophical Transactions of the Royal Society A* 370: 2123–2142.

Coulthard, T.J., Macklin, M.G. and Kirkby, M.J. 2002. A cellular model of Holocene upland river basin and alluvial fan evolution. *Earth Surface Processes and Landforms* 27: 269–288.

Coulthard T.J., Lewin, J. and Macklin, M.G. 2005. Modelling differential catchment response to environmental change. *Geomorphology* 69: 222–241.

Coulthard, T.J., Hicks, M.D. and Van de Wiel, M.J. 2007. Cellular modelling of river catchments and reaches: advantages, limitations and prospects. *Geomorphology* 90: 192–207.

Couper P., Scott, T. and Maddock, I. 2002. Insights into river bank erosion processes derived from analysis of negative erosion-pin recordings: observations from three UK studies. *Earth Surface Processes and Landforms* 27: 59–79.

Crosato, A. and Saleh, M.S. 2010. Numerical study on the effects of floodplain vegetation on river planform style. *Earth Surface Processes and Landforms* 36: 711–720.

Dapporto, S., Rinaldi, M. and Casagli, N. 2001. Failure mechanisms and pore water pressure conditions: analysis of a riverbank along the Arno River (Central Italy). *Engineering Geology* 61: 221–242.

Daraio, J.A., Weber, L.J., Newton, T.J. and Nestler, J.M. 2010. A methodological framework for integrating computational fluid dynamics and ecological models applied to juvenile freshwater mussel dispersal in the Upper Mississippi River. *Ecological Modelling* 221: 201–214.

Darby, S. and Sear, D. 2008. *River Restoration: Managing the Uncertainty in Restoring Physical Habitat*, Chichester: John Wiley & Sons.

Darby, S.E. and Thorne, C.R. 1996a. Modelling the sensitivity of channel adjustments in destabilized sand-bed rivers. *Earth Surface Processes and Landforms* 21: 1109–1125.

Darby, S.E. and Thorne, C.R. 1996b. Numerical simulation of widening and bed deformation of straight sand-bed rivers. I. Model development. *Journal of Hydraulic Engineering* 122: 184–193.

Darby, S.E., Gessler, D. and Thorne, C.R. 2000. Computer program for stability analysis of steep, cohesive riverbanks. *Earth Surface Processes and Landforms* 25: 175–190.

Darby, S.E., Alabyan, A.M. and Van de Wiel, M.J. 2002. Numerical simulation of bank erosion and channel migration in meandering rivers. *Water Resources Research* 38: 1163.

Darby, S.E., Rinaldi, M. and Dapporto, S. 2007. Coupled simulations of fluvial erosion and mass wasting for cohesive river banks. *Journal of Geophysical Research* 112: F03022.

Davies, T.R.H. and Sutherland, A.J. 1983. Extremal hypotheses for river behaviour. *Water Resources Research* 19: 141–148.

Davis, W.M., 1899. The geographical cycle. *Geographical Journal* 14: 481–504.

Davy, B.W. and Davies, T.R.H. 1979. Entropy concepts in fluvial geomorphology: a reevaluation. *Water Resources Research* 15: 103–106.

De Boer, D.H. 1992. Hierarchies and spatial scale in process geomorphology: a review. *Geomorphology* 4: 303–318.

De Rose, R. and Basher, L. 2011. Measurement of river bank and cliff erosion from sequential LIDAR and historical aerial photography. *Geomorphology* 126: 132–147.

Docker, B.B. and Hubble, T.C.T. 2008. Quantifying root-reinforcement of river bank soils by four Australian tree species. *Geomorphology* 100: 401–418.

Downs, P.W. and Brookes, A. 1994. Developing a standard geomorphological approach for the appraisal of river projects. In: Kirby, C. and White, W.R., eds., *Integrated River Basin Development*, Chichester: John Wiley & Sons, pp. 299–310.

Downs, P.W. and Priestnall, G. 1999. System design for catchment-scale approaches to studying river channel adjustments using a GIS. *International Journal of Geographical Information Science* 13(3): 247–266.

Duan, J.G. and Julien, P.Y. 2005. Numerical simulation of the inception of channel meandering. *Earth Surface Processes and Landforms* 30: 1093–1110.

Duan, J.G. and Julien, P.Y. 2010. Numerical simulation of meandering evolution. *Journal of Hydrology* 391: 34–46.

Dutta, D., Alam, J., Umeda, K., Hayashi, M. and Hironaka, S. 2007. A two-dimensional hydrodynamic model for flood inundation simulation: a case study in the lower Mekong river basin. *Hydrological Processes* 21: 1223–1237.

Edwards, B.F. and Smith, D.H. 2001. Critical wavelength for river meandering. *Physical Review E* 63: 045304.

Edwards, B.F. and Smith, D.H. 2002. River meandering dynamics. *Physical Review E* 65: 046303.

Einstein, A. 1926. The cause of the formation of meanders in the courses of rivers and of the so-called Baer's law. *Die Naturwissenschaften* 14: 223–224 (in German).

Fang, H.-W. and Wang, G.-Q. 2000. Three-dimensional mathematical model of suspended-sediment transport. *Journal of Hydraulic Engineering* 126: 578–592.

Farajalla, N.S. and Vieux, B.E. 1995. Capturing the essential spatial variability in distributed hydrological modelling: infiltration parameters. *Hydrological Processes* 9: 55–68.

Fellenius, W. 1936. Calculation of the stability of earth dams. In: *Transactions of the Second International Congress on Large Dams*, vol. 4, pp. 445–462.

Ferguson, R.I. 1986. Hydraulics and hydraulic geometry. *Progress in Physical Geography* 10: 1–31.

Ferguson, R.I., Parsons, D.R., Lane, S.N. and Hardy, R.J. 2003. Flow in meander bends with recirculation at the inner bank. *Water Resources Research* 39: 1322.

Fischer-Antze, T., Rüther, N., Olsen, N.R.B. and Gutknecht, D. 2009. Three-dimensional (3D) modeling of non-uniform sediment transport in a channel bend with unsteady flow. *Journal of Hydraulic Research* 47: 670–675.

Fonstad, M. and Marcus, W.A. 2003. Self-organized criticality in riverbank systems. *Annals of the Association of American Geographers* 93: 281–296.

Frascati, A. and Lanzoni, S. 2009. Morphodynamic regime and long-term evolution of meandering rivers. *Journal of Geophysical Research* 114: F02002.

Frascati, A. and Lanzoni, S. 2010. Long-term river meandering as a part of chaotic dynamics? A contribution from mathematical modelling. *Earth Surface Processes and Landforms* 35: 791–802.

Frey, P. and Church, M. 2011. Bedload: a granular phenomenon. *Earth Surface Processes and Landforms* 36: 58–69.

Friedkin, J.F. 1945. *A Laboratory Study of the Meandering of Alluvial Rivers*, Vicksburg, MS: US Army Corps of Engineers.

Gardner, J.T. and Ashmore, P.E. 2011. Geometry and grain-size characteristics of the basal surface of a braided river deposit. *Geology* 39: 247–250.

Geist, D.R. and Dauble, D.D. 1998. Redd site selection and spawning habitat use by fall chinook salmon: the importance of geomorphic features in large rivers. *Environmental Management* 22: 655–669.

Gervasio, P., Ovtchinnikov, E. and Quarteroni, A. 1997. The spectral projection decomposition method for elliptic equations in two dimensions. *SIAM Journal on Numerical Analysis* 34: 1616–1639.

Glock, W.S. 1931. The development of drainage systems: a synoptic view. *Geographical Review* 21: 475–482.

Glover, R.E. and Florey, Q.L. 1951. *Stable Channel Profiles*, Denver, CO: US Bureau of Reclamation.

Gomez, B. and Church, M. 1989. An assessment of bed load sediment transport formulae for gravel bed rivers. *Water Resources Research* 25: 1161–1186.

Gran, K. and Paola, C. 2001. Riparian vegetation controls on braided stream dynamics. *Water Resources Research* 37: 3275–3283.

Grant, G. E., O'Connor, J.E. and Wolman, M.G. 2013. A river runs through it: conceptual models in fluvial geomorphology. In: Shroder, J. and Wohl, E., eds., *Treatise on Geomorphology, Volumr 9, Fluvial Geomorphology*, San Diego, CA: Academic Press, pp. 6–21.

Griffiths, G.A. 1984. Extremal hypotheses for river regime: an illusion of progress. *Water Resources Research* 20: 113–118.

Guo, Q.C. and Jin, Y.C. 2002. Modeling nonuniform suspended sediment transport in alluvial rivers. *Journal of Hydraulic Engineering* 128: 839–847.

Haff, P.K. 1996. Limitations on predictive modelling in geomorphology. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 337–358.

Haltigin, T.W., Biron, P.M. and Lapointe, M.P. 2007. Three-dimensional numerical simulation of flow around stream deflectors: the effect of obstruction angle and length. *Journal of Hydraulic Research* 45(2): 227–238.

Hankin, B.C. and Beven, K.J. 1998. Modeling dispersion in complex open channel flows: fuzzy calibration. *Stochastic Hydrology and Hydraulics* 12: 397–412.

Hardy, R.J., Bates, P.D. and Anderson, M.G. 1999. The importance of spatial resolution in hydraulic models for floodplain environments. *Journal of Hydrology* 216: 124–136.

Hardy, R.J., Bates, P.D. and Anderson, M.G. 2000. Modelling suspended sediment deposition on a fluvial floodplain using a two-dimensional finite element model. *Journal of Hydrology* 229: 202–218.

Hasbargen, L.E. and Paola, C. 2000. Landscape instability in an experimental drainage basin. *Geology* 28: 1067–1070.

Hasegawa, K. 1977. Computer simulation of meandering channels. *Proceedings of the Hokkaido Branch, Japan Society of Civil Engineering* 197–202 (in Japanese).

Hervouet, J.-M. 2000. A high resolution 2-D dam-break model using parallelization. *Hydrological Processes* 14: 2211–2230.

Hervouet J.-M. and Van Haren, L. 1996. Recent advances in numerical methods for fluid flows. In: Anderson, M.G., Walling, D.E. and Bates, P.D., eds., *Floodplain Processes*, Chichester: John Wiley & Sons, pp. 183–214.

Hey, R.D. 1978. Determinate hydraulic geometry of river channels. *Journal of the Hydraulics Division, American Society of Civil Engineers* 104: 869–885.

Hey, R.D. and Heritage, G.L. 1988. Dimensional and dimensionless regime equations for gravel-bed rivers. In: White, W.R., ed., *River Regime*, Chichester: John Wiley & Sons, pp. 1–8.

Hickin, E.J. 1977. Hydraulic factors controlling channel migration. In: Davidson-Arnott, R. and Nickling, W., eds., *Research in Fluvial Systems*, Norwich: Geo-Books, pp. 59–72.

Hickin, E.J. 1978. Mean flow structure in meanders of the Squamish River, British Columbia. *Canadian Journal of Earth Sciences* 15: 1833–1849.

Hickin, E.J. and Nanson, G.C. 1975. The character of channel migration on the Beatton River, Northeast British Columbia, Canada. *Bulletin of the Geological Society of America* 86: 487–494.

Hodskinson, A. and Ferguson, R.I. 1998. Numerical modelling of separated flow in river bends: model testing and experimental investigation of geometric controls on the extent of flow separation at the concave bank. *Hydrological Processes* 12: 1323–1338.

Hooke, J.M. 1997. Styles of channel change. In: Thorne, C.R., Hey, R.D. and Newson, M.D., eds., *Applied Fluvial Geomorphology for River Engineering and Management*, Chichester: John Wiley & Sons, pp. 237–268.

Hooke, J.M. 2007. Complexity, self-organisation and variation in behaviour in meandering rivers. *Geomorphology* 91: 236–258.

Horritt, M.S. and Bates, P.D. 2001. Predicting floodplain inundation: raster-based modelling versus the finite element approach. *Hydrological Processes* 15: 825–842.

Horritt, M.S. and Bates, P.D. 2002. Evaluation of 1D and 2D numerical models for predicting flood inundation. *Journal of Hydrology* 268: 825–842.

Howard, A.D. 1992. Modelling channel migration and floodplain development in meandering streams. In: Carling, P.A. and Petts, G.E., eds., *Lowland Floodplain Rivers*, Chichester: John Wiley & Sons, pp. 1–42.

Howard, A.D. 1994. A detachment-limited model of drainage basin evolution. *Water Resources Research* 30: 2261–2285.

Howard, A.D. 1996. Modelling channel evolution and floodplain morphology. In: Anderson, M.G., Walling, D.E. and Bates, P.D., eds., *Floodplain Processes*, Chichester: John Wiley & Sons, pp.15–62.

Howard, A.D. and Knutson, T.R. 1984. Sufficient conditions for river meandering: a simulation approach. *Water Resources Research* 20: 1659–1667.

Howard, A.D., Dietrich, W.E. and Seidl, M.A. 1994. Modeling fluvial erosion on regional to continental scales. *Journal of Geophysical Research* 99: 13971–13986.

Huang, H.Q. and Nanson, G.C. 2002. A stability criterion inherent in laws governing alluvial channel flow. *Earth Surface Processes and Landforms* 27: 929–944.

Ikeda, S. 1981. Self-formed straight channels in sandy beds. *Journal of the Hydraulics Division, American Society of Civil Engineers* 107: 389–406.

Ikeda, S., Parker, G. and Sawai, K. 1981. Bend theory of river meanders. 1: Linear development. *Journal of Fluid Mechanics* 112: 363–377.

Jia, Y., Scott, S., Xu, Y., Huang, Y. and Wang, S.S.Y. 2005. Three-dimensional numerical simulation and analysis of flows around a submerged weir in a channel bendway. *Journal of Hydraulic Engineering* 131(8): 682–693.

Johannesson, H. and Parker, G. 1989. Secondary flow in mildly sinuous channels. *Journal of Hydraulic Engineering* 115: 289–308.

Johnson, P.A., Gleason, G.L. and Hey, R.D. 1999. Rapid assessment of channel stability in vicinity of road crossing. *Journal of Hydraulic Engineering* 125: 645–651.

Kanaroglou, P.S. 1996. On spurious correlation in geographical problems. *Canadian Geographer* 40: 194–202.

Kassem, A.A. and Chaudhry, M.H. 2002. Numerical evolution of bed evolution in channel bends. *Journal of Hydraulic Engineering* 128: 507–552.

Keller, E.A. 1972. Development of alluvial stream channels: a 5 stage model. *Bulletin of the Geological Society of America* 83: 1531–1536.

Kellerhals, R., Church, M. and Bray, D.I. 1976. Classification and analysis of river processes. *Journal of the Hydraulics Division. American Society of Civil Engineers* 102: 813–829.

Kirkby, M.J. 1990. The landscape viewed through models. *Zeitschrift für Geomorphologie* 79: 63–81.

Kirkby, M.J. 1996. A role for theoretical models in geomorphology? In: Rhoads, B.L. and Thorn, C.E., eds, *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 257–272.

Kolditz, O. 2002. Finite volume method. In: *Computational Methods in Environmental Fluid Mechanics*, New York: Springer, pp. 173–190.

Konikow, L.F. and Bredehoeft, J.D. 1992. Ground-water models cannot be validated. *Advances in Water Resources* 15: 75–83.

Kovacs, A. and Parker, G. 1994. A new vectorial bedload formulation and its application to the time evolution of straight river channels. *Journal of Fluid Mechanics* 267: 153–183.

Kresic, N. 2007. Groundwater modeling. In: *Hydrogeology and Groundwater Modeling*, 2nd edn., Boca Raton, FL: CRC Press, pp. 499–547.

Lajeunesse, E., Malverti, L. and Charru, F. 2010. Bed load transport in turbulent flow at the grain scale: experiments and modeling. *Journal of Geophysical Research* 115: F04001.

Lam, L and Fredlund, D.G. 1993. A general limit equilibrium model for three-dimensional slope stability analysis. *Canadian Geotechnical Journal* 30: 905–919.

Lancaster, S.T. and Bras, R.L. 2002. A simple model of river meandering and its comparison to natural channels. *Hydrological Processes* 16: 1–26.

Lane, E.W. 1955. Design of stable channels. *Transactions of the American Society of Civil Engineers* 120: 1234–1260.

Lane, S.N. 1998. Hydraulic modelling in hydrology and geomorphology: a review of high resolution approaches. *Hydrological Processes* 12: 1131–1150.

Lane, S.N. and Richards, K.S. 1998. High resolution, two-dimensional spatial modelling of flow processes in a multi-thread channel. *Hydrological Processes* 12: 1279–1298.

Lane, S.N., Tayefi, V., Reid, S.C., Yu, D. and Hardy, R.J. 2007. Interactions between sediment delivery, channel change, climate change and flood risk in a temperate upland environment. *Earth Surface Processes and Landforms* 32: 429–446.

Langendoen, E.J. and Simon, A. 2008. Modeling the evolution of incised streams. II. Streambank erosion. *Journal of Hydraulic Engineering* 134: 905–915.

Langendoen, E.J., Lowrance, R.R. and Simon, A. 2009. Assessing the impact of riparian processes on streambank stability. *Ecohydrology* 2: 360–369.

Larned, S.T., Schmidt, J., Datry, T., Konrad, C.P., Dumas, J.K. and Diettrich, J.C. 2011. Longitudinal river ecohydrology: flow variation down the lengths of alluvial rivers. *Ecohydrology* 4(4):532–548.

Lee, J.-S. and Julien, P.Y. 2006. Downstream hydraulic geometry of alluvial channels. *Journal of Hydraulic Engineering* 132(12): 1347–1352.

Leeder, M.R. and Bridges, P.H. 1975. Flow separation in meander bends. *Nature* 253: 338–339.

Legleiter, C. and Roberts, D. 2009. A forward image model for passive optical remote sensing of river bathymetry. *Remote Sensing of Environment* 113, 1025–1045.

Legleiter, C.J., Roberts, D.A. and Lawrence, R.L. 2009. Spectrally based remote sensing of river bathymetry. *Earth Surface Processes and Landforms* 34: 1039–1059.

Leopold, L.B. and Maddock, T. 1953. *The Hydraulic Geometry of Stream Channels and Some Physiographic Implications*, US Geological Survey Professional Paper 252, Washington, DC: US Government Printing Office.

Leyland, J. and Darby, S.E. 2009. Effects of Holocene climate and sea-level changes on coastal gully evolution: insights from numerical modelling. *Earth Surface Processes and Landforms* 34: 1878–1893.

Li, L. and Wang, S.S.Y. 1993. Numerical modelling of alluvial stream bank erosion. In: Wang, S.S.Y., ed., *Advances in Hydro-Science and Engineering*, Oxford, MS: University of Mississippi, pp. 2085–2090.

Liao, C.B., Wu, M.S. and Liang, S.J. 2007. Numerical simulation of a dam break for an actual river terrain environment. *Hydrological Processes* 21: 447–460.

Lien, H.C., Hsieh, T.Y., Yang, J.C. and Yeh, K.C. 1999. Bend-flow simulation using 2D depth-averaged model. *Journal of Hydraulic Engineering* 125: 1097–1108.

Lohnes, R. 1991. A method for estimating land loss associated with stream channel degradation. *Engineering Geology* 31: 115–130.

Luchi, R., Zolezzi, G. and Tubino, M. 2010. Modelling mid-channel bars in meandering channels. *Earth Surface Processes and Landforms* 35: 902–917.

Luchi, R., Zolezzi, G. and Tubino, M. 2011. Bend theory of river meanders with spatial width variations. *Journal of Fluid Mechanics* 681: 311–339.

Luppi, L., Rinaldi, M., Teruggi, L.B., Darby, S.E. and Nardi, L. 2009. Monitoring and numerical modelling of riverbank erosion processes: a case study along the Cecina River (central Italy). *Earth Surface Processes and Landforms* 34: 530–546.

Maddux, T.B., Nelson, J.M. and McLean, S.R. 2003. Turbulent flow over three-dimensional dunes: 1. Free surface and flow response. *Journal of Geophysical Research* 108: 6009.

Mao, L., Cooper, J.R. and Frostick, L.E. 2011. Grain size and topographical differences between static and mobile armour layers. *Earth Surface Processes and Landforms* 36: 1321–1334.

Markham, A.J. and Thorne, C.R. 1992. Geomorphology of gravel-bed river bends. In: Billi, P., Hey, R.D., Thorne, C.R. and Tacconi, P., eds., *Dynamics of Gravel-Bed Rivers*, Chichester: John Wiley & Sons, pp. 433–456.

Martin, J., Sheets, B., Paola, C. and Hoyal, D. 2009. Influence of steady base-level rise on channel mobility, shoreline migration and scaling properties of a cohesive experimental delta. *Journal of Geophysical Research* 115: F03017.

Meakin, P., Sun, T., Jøssang, T. and Schwartz, K. 1996. A simulation model for meandering rivers and their associated sedimentary environments. *Physica A* 233: 606–618.

Mekonnen, M.A., Secretan, Y. and Biron, P.M. 2010. Two-dimensional morphodynamic model for prediction of changes in river bed level and sedimentary profile. In: *Proceedings of the 8th International Symposium on Ecohydraulics*, Seoul, Korea, pp. 750–757.

Melville, B.W. and Sutherland, A.J. 1988. Design method for local scour at bridge piers. *Journal of Hydraulic Engineering* 114: 1210–1226.

Michaelides, K. and Wainwright, J. 2008. Internal testing of a numerical model of hillslope–channel coupling using laboratory flume experiments. *Hydrological Processes* 22: 2274–2291.

Midgley, T.L., Fox, G.A. and Heeren, D.M. 2012. Evaluation of the bank stability and toe erosion model (BSTEM) for predicting lateral retreat on composite streambanks. *Geomorphology* 145–146: 107–114.

Milhous, R.T. 1998a. Numerical modelling of flushing flows in gravel-bed rivers. In: Klingeman, P.C., Beschta, R.L., Komar, P.D. and Bradley, J.B., eds', *Gravel-bed Rivers in the Environment*, Highlands Ranch, CO: Water Resources Publications, pp. 579–608.

Milhous, R.T. 1998b. Modelling of instream flow needs: the link between sediment and aquatic habitat. *Regulated Rivers – Research and Management* 14: 79–94.

Millar, R.G. and Quick, M.C. 1993. Effect of bank stability on geometry of gravel rivers. *Journal of Hydraulic Engineering* 119: 1343–1363.

Millar, R.G. and Quick, M.C. 1998. Stable width and depth of gravel-bed rivers with cohesive banks. *Journal of Hydraulic Engineering*, 124: 1005–1013.

Miller, T.K. 1984. A system model of stream channel shape and size. *Bulletin of the Geological Society of America* 95: 237–241.

Montgomery, D.R. and Buffington, J.M. 1997. Channel-reach morphology in mountain drainage basins. *Bulletin of the Geological Society of America* 109: 596–611.

Morgenstern, N.R. and Price, V.E. 1965. The analysis of the stability of general slip surfaces. *Géotechnique* 15(1): 79–93.

Mosselman, E. 1995. A review of mathematical models of river planform changes. *Earth Surface Processes and Landforms* 20: 661–670.

Mosselman, E. 1998. Morphological modelling of rivers with erodible banks. *Hydrological Processes* 12: 1357–1370.

Murray, A.B. 2003. Contrasting the goals, strategies and predictions associated with simplified numerical models and detailed simulations. In: Wilcock, P.R. and Iverson, R.M., eds., *Prediction in Geomorphology*, Washington, DC: American Geophysical Union, pp. 151–165.

Murray, A.B. 2007. Reducing model complexity for explanation and prediction. *Geomorphology* 90: 178–191.

Murray, A.B. and Paola, C. 1994. A cellular model of braided rivers. *Nature* 371: 54–56.

Murray, A.B. and Paola, C. 1997. Properties of a cellular braided stream model. *Earth Surface Processes and Landforms* 22: 1001–1025.

Murray, A.B. and Paola, C. 2003. Modelling the effect of vegetation on channel pattern in bedload rivers. *Earth Surface Processes and Landforms* 28: 131–143.

Nagata, N., Hosoda, T. and Muramoto, Y. 2000. Numerical analysis of river channel processes with bank erosion. *Journal of Hydraulic Engineering* 126: 243–252.

Nanson, G.C. and Huang, H.Q. 2008. Least action principle, equilibrium states, iterative adjustment and the stability of alluvial channels. *Earth Surface Processes and Landforms* 33: 923–942.

Nicholas, A.P. 2009. Reduced-complexity flow routing models for sinuous single-thread channels: intercomparison with a physically-based shallow-water equation model. *Earth Surface Processes and Landforms* 34: 641–653.

Nicholas, A.P. 2010. Reduced-complexity modeling of free bar morphodynamics in alluvial channels. *Journal of Geophysical Research* 115: F04021.

Nicholas, A.P. and Quine, T.A. 2007. Crossing the divide: representation of channels and processes in reduced-complexity river models at reach and landscape scales. *Geomorphology* 90: 318–339.

Nicholas, A.P. and Sambrook Smith, G.H. 1999. Numerical simulation of three-dimensional flow hydraulics in a braided channel. *Hydrological Processes* 13: 913–929.

Nicholas, A.P. and Walling, D.E. 1998. Numerical modelling of floodplain hydraulics and suspended sediment transport and deposition. *Hydrological Processes* 12: 1339–1355.

Notebaert, B., Verstraeten, G., Govers, G. and Poesen, J. 2009. Qualitative and quantitative applications of LiDAR imagery in fluvial geomorphology. *Earth Surface Processes and Landforms* 34: 217–231.

Odgaard, A.J. 1989. River meander model. I. Development. *Journal of Hydraulic Engineering* 115: 1433–1450.

Olsen, N.R.B. 2003. Three-dimensional CFD modeling of self-forming meandering channel. *Journal of Hydraulic Engineering* 129: 366–372.

Olsen, N.R.B. and Kjellesvig, H.M. 1998. Three-dimensional numerical flow modeling for estimation of maximum local scour depth. *Journal of Hydraulic Research* 36: 579–590.

Olsen, N.R.B. and Stokseth, S. 1995. Three-dimensional numerical modelling of water flow in a river with large bed roughness. *Journal of Hydraulic Research* 33: 571–581.

O'Neill, M.P., Schmidt, J.C., Dobrowolski, J.P., Hawkins, C.P. and Neale, C.M.U. 1997. Identifying sites for riparian wetland restoration: application of a model to the Upper Arkansas River basin. *Restoration Ecology* 5: 85–102.

Oreskes, N., Shrader-Frechette, K. and Belitz, K. 1994. Verification, validation and confirmation of numerical models in the earth sciences. *Science* 263: 641–646.

Osman, A.M. and Thorne, C.R. 1988. Riverbank stability analysis. I. Theory. *Journal of Hydraulic Engineering* 124: 134–150.

Ouchi, S. 2011. Effects of uplift on the development of experimental erosion landform generated by artificial rainfall. *Geomorphology* 127: 88–98.

Paik, K. and Kumar, P. 2010. Optimality approaches to describe characteristic fluvial patterns on landscapes. *Philosophical Transactions of the Royal Society B* 365: 1387–1395.

Paine, A.D.M. 1985. 'Ergodic' reasoning in geomorphology. *Progress in Physical Geography* 9: 1–15.

Parker, G., Shimizu, Y., Wilkerson, G.V., Eke, E.C., Abad, J.D., Lauer, J.W., Paola, C., Dietrich, W.E. and Voller, V.R. 2011. A new framework for modeling the migration of meandering rivers. *Earth Surface Processes and Landforms* 36: 70–86.

Patton, P.C. and Baker, V.R. 1977. Geomorphic response of central Texas stream channels to catastrophic rainfall and runoff. In: Doehring, D.O., ed., *Geomorphology in Arid and Semi-Arid Regions*, Publications in Geomorphology, Binghamton, NY: State University of New York, pp. 189–217.

Peakall, J., Ashworth, P.J. and Best, J. 1996. Physical modelling in fluvial geomorphology: principles, applications and unresolved issues. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 221–253.

Perucca, E., Camporeale, C. and Ridolfi, L. 2007. Significance of the riparian vegetation dynamics on meandering river morphodynamics. *Water Resources Research* 43: W03430.

Pinelli, A., Vacca, A. and Quarteroni, A. 1997. A spectral multidomain method for the numerical simulation of turbulent flows. *Journal of Computational Physics* 136: 546–558.

Pizzuto, J.E. 1990. Numerical simulation of gravel river widening. *Water Resources Research* 26: 1971–1980.

Pizzuto, J.E. 1992. The morphology of graded gravel rivers: a network perspective. *Geomorphology* 5: 457–474.

Pizzuto J., O'Neil, M. and Stotts, S. 2010. On the retreat of forested, cohesive riverbanks. *Geomorphology* 116: 341–352.

Pollen, N. 2007. Temporal and spatial variability in root reinforcement of streambanks: accounting for soil shear strength and moisture. *Catena* 69: 197–205.

Pollen, N., Simon, A. and Collison, A. 2004. Advances in assessing the mechanical and hydrologic effects of riparian vegetation on streambank stability. In: Bennett, S.J. and Simon, A., eds., *Riparian Vegetation and Fluvial Geomorphology*, Washington, DC: American Geophysical Union, pp. 125–139.

Pollen-Bankhead, N. and Simon, A. 2009. Enhanced application of root-reinforcement algorithms for bank-stability modeling. *Earth Surface Processes and Landforms* 34: 471–480.

Pollen-Bankhead, N. and Simon, A. 2010. Hydrologic and hydraulic effects of riparian root networks on streambank stability: is mechanical root-reinforcement the whole story? *Geomorphology* 116: 353–362.

Posner, A. and Duan, J.G. 2012. Simulating river meandering processes using stochastic bank erosion coefficient. *Geomorphology:* 163–164: 26–36.

Postma, G., Kleinhans, M.G., Meijer, P.T. and Eggenhuisen, J.T. 2008. Sediment transport in analogue flume models compared with real-world sedimentary systems: a new look at scaling evolution of sedimentary systems in a flume. *Sedimentology* 55: 1541–1557.

Refsgaard, J. 2000. Towards a formal approach to calibration and validation of models using spatial data. In: Grayson, R. and Blöschl, G., eds., *Spatial Patterns in Catchment Hydrology: Observations and Modelling*, Cambridge: Cambridge University Press, pp. 329–354.

Resop J. and Hession, W. 2010. Terrestrial laser scanning for monitoring streambank retreat: comparison with traditional surveying techniques. *Journal of Hydraulic Engineering* 136(10): 794–798.

Reubens, B., Poesen, J., Danjon, F., Geudens, G. and Muys, B. 2007. The role of fine and coarse roots in shallow slope stability and soil erosion control with a focus on root system architecture: a review. *Trees – Structure and Function* 21: 385–402.

Rhoads, B.L. 1992. Statistical models of fluvial systems. *Geomorphology* 5: 433–455.

Rice, J.R. 1983. *Numerical Methods, Software and Analysis*, ISML Reference Edition, New York: McGraw-Hill.

Rice, S.P., Lancaster, J. and Kemp, P. 2009. Experimentation at the interface of fluvial geomorphology, stream ecology and hydraulic engineering and the development of an effective, interdisciplinary river science. *Earth Surface Processes and Landforms* 35: 64–77.

Richards, K. 1996. Samples and cases: generalisation and explanation in geomorphology. In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 171–190.

Rinaldi, M. and Casagli, N. 1999. Stability of streambanks formed in partially saturated soils and effects of negative pore water pressures: the Sieve River, Italy. *Geomorphology* 26: 253–277.

Rinaldi, M., Casagli, N., Dapporto, S. and Gargini, A. 2004. Monitoring and modelling of pore water pressure changes and riverbank stability during flow events. *Earth Surface Processes and Landforms* 29: 237–254.

Rinaldi, M., Mengoni B., Luppi L., Darby S.E. and Mosselman E. 2008. Numerical simulation of hydrodynamics and bank erosion in a river bend. *Water Resources Research* 44: W09428.

Rojstaczer, S.A. 1994. The limitations of ground water models. *Journal of Geological Education* 42: 362–368.

Rosenfeld, J.S., Campbell, K., Leung, E.S., Bernhardt, J. and Post, J. 2011. Habitat effects on depth and velocity frequency distributions: implications for modeling hydraulic variation and fish habitat suitability in streams. *Geomorphology* 130: 127–135.

Rüther, N. and Olsen, N.R.B. 2005. Three-dimensional modeling of sediment transport in a narrow 90° channel bend. *Journal of Hydraulic Engineering* 131: 917–920.

Sack, D. 1992. New wine in old bottles: the historiography of a paradigm change. *Geomorphology* 5: 251–263.

Samadi, A., Amiri-Tokaldany, E. and Darby, S.E. 2009. Identifying the effects of parameter uncertainty on the reliability of riverbank stability modelling. *Geomorphology* 106: 219–230.

Samadi, A, Amiri-Tokaldany, E., Davoudi, M.H. and Darby, S.E. 2011. Identifying the effects of parameter uncertainty on the reliability of modeling the stability of overhanging, multi-layered, river banks. *Geomorphology* 134: 483–498.

Schumm, S.A. 1991. *To Interpret the Earth: Ten Ways to be Wrong*, Cambridge: Cambridge University Press.

Schumm, S.A., Harvey, M.D. and Watson, C.C. 1984. *Incised River Channels: Morphology, Dynamics and Control*, Highlands Ranch, CO: Water Resources Publications.

Schumm, S.A., Mosley, M.P. and Weaver, W.E. 1987. *Experimental Fluvial Geomorphology*, New York: John Wiley & Sons.

Sear, D., Newson, M., Hill, C., Old, J. and Branson, J. 2009. A method for applying fluvial geomorphology in support of catchment-scale river restoration planning. *Aquatic Conservation – Marine and Freshwater Ecosystems* 19: 506–519.

Seminara, G. 2006. Meanders. *Journal of Fluid Mechanics* 554: 271–297.

Seminara, G. and Tubino, M. 1989. Alternate bars and meandering: free, forced and mixed interactions. In: Ikeda, S. and Parker, G., eds., *River Meandering*, Washington, DC: American Geophysical Union, pp. 267–320.

Seminara, G., Zolezzi, G., Tubino, M. and Zardi, D. 2001. Downstream and upstream influence in river meandering. Part 2. Planimetric development. *Journal of Fluid Mechanics* 438: 213–230.

Sewell, G. 1988. *The Numerical Solution of Ordinary and Partial Differential Equations*, Boston, MA: Academic Press.

Sheets, B.A., Hickson, T.A. and Paola, C. 2002. Assembling the stratigraphic record: depositional patterns and time-scales in an experimental alluvial basin. *Basin Research* 14: 287–301.

Shen, H.W., ed. 1991. *Movable Bed Physical Models*, NATO ISI Series, Series C: Mathematical and Physical Sciences, vol. 312, Dordrecht: Kluwer.

Shields, F.D. 1996. Hydrologic and hydraulic stability. In: Brookes A. and Shields, F.D. Jr,, eds., *River Channel Restoration*, Chichester: John Wiley & Sons, pp. 75–101.

Shields, F.D., Knight, S.S. and Cooper, C.M. 1998. Rehabilitation of aquatic habitats in warmwater streams damaged by channel incision in Mississippi. *Hydrobiologia* 382: 63–86.

Simon, A. 1989. A model of channel response in disturbed alluvial channels. *Earth Surface Processes and Landforms* 14: 11–26.

Simon, A. 1992. Energy, time and channel evolution in catastrophically disturbed fluvial systems. *Geomorphology* 5: 345–372.

Simon, A and Collison, A.J.C. 2002. Quantifying the mechanical and hydrologic effects of riparian vegetation on streambank stability. *Earth Surface Processes and Landforms* 27: 527–546.

Simon, A. and Downs, P.W. 1995. An interdisciplinary approach to evaluation of potential instability in alluvial channels. *Geomorphology* 12: 215–232.

Simon, A., Wolfe, W.J. and Molinas, A. 1991. Mass wasting algorithms in an alluvial channel model. Presented at the 5th Federal Interagency Sedimentation Conference, 22–29 September 1991, Las Vegas, NV.

Simon, A., Curini, A., Darby, S.E. and Langendoen, E.J. 2000. Bank and near-bank processes in an incised channel. *Geomorphology* 35: 193–217.

Simon, A., Pollen, N. and Langendoen, E.J. 2006. Influence of two woody riparian species on critical conditions for streambank stability: Upper Truckee River, California. *Journal of the American Water Resources Association* 42: 99–113.

Slingerland, R. and Smith, N.D. 1998. Necessary conditions for a meandering river avulsion. *Geology* 26: 435–438.

Smith, J.D. and McLean, S.R. 1984. A model for flow in meandering streams. *Water Resources Research* 20: 1301–1315.

Spencer, E. 1967. A method of analysis of the stability of embankments assuming parallel inter-slice forces. *Géotechnique* 171: 11–26.

Stewardson, M. and Rutherfurd, I. 2008. Conceptual and mathematical modelling in river restoration: do we have unreasonable confidence? In: Darby, S.E. and Sear, D.A., eds., *River Restoration: Managing the Uncertainty in Restoring Physical Habitat*, Chichester: John Wiley, pp. 61–78.

Stoesser, T., Braun, C., García-Villalba, M. and Rodi, W. 2008. Turbulence structures in flow over two-dimensional dunes. *Journal of Hydraulic Engineering* 134: 42–55.

Stølum, H.-H. 1996. River meandering as a self-organization process. *Science* 271: 1710–1713.

Stølum, H.-H. 1998. Planform geometry and dynamics of meandering rivers. *Geological Society of America Bulletin* 110: 1485–1498.

Strahler, A.N. 1980. Systems theory in physical geography. *Physical Geography* 1: 1–27.

Sun, T., Meakin, P., Jøssang, T. and Schwarz, K. 1996. A simulation model for meandering rivers. *Water Resources Research* 32: 2937–2954.

Sun, T., Meakin, P. and Jøssang, T. 2001. A computer model for meandering rivers with multiple bedload sediment sizes. 1. Theory. *Water Resources Research* 37: 2227–2241.

Tal, M., Gran, K., Murray, A.B., Paola, C. and Hicks, D.M. 2004. Riparian vegetation as a primary control on channel characteristics in multi-thread rivers. In: Bennett, S.J. and Simon, A., eds., *Riparian Vegetation and Fluvial Geomorphology*, Washington, DC: American Geophysical Union, pp. 43–58.

Temme, A.J.A.M., Baartman, J.E.M. and Schoorl, J.M. 2009. Can uncertain landscape evolution models discriminate between landscape responses to stable and changing future climate? A millennial-scale test. *Global and Planetary Change* 69: 48–58.

Tetzlaff, D.M. 1989. Limits to the predictive stability of dynamic models that simulate clastic sedimentation. In: Cross, T.A., ed., *Quantitative Dynamic Stratigraphy*, Englewood Cliffs, NJ: Prentice-Hall, pp. 55–65.

Tetzlaff, D.M. and Harbaugh, J.W. 1989. *Simulating Clastic Sedimentation*, New York: Van Nostrand Reinhold.

Thompson, A. 1986. Secondary flows and the pool–riffle unit: a case study of the processes of meander development. *Earth Surface Processes and Landforms* 11: 631–641.

Thomson, J., 1876. On the origins and winding of rivers in alluvial plains. *Proceedings of the Royal Society of London* 25: 5–8.

Thonon, I., De Jong, K., Van der Perk, M. and Middelkoop, H. 2007. Modelling floodplain sedimentation using particle tracking. *Hydrological Processes* 21: 1402–1412.

Thorne, C.R. 1995. Editorial – Geomorphology at work. *Earth Surface Processes and Landforms* 20: 583–584.

Thorne, C.R. and Tovey, N.K. 1981. Stability of composite river banks. *Earth Surface Processes and Landforms* 6: 469–484.

Thorne, C.R., Allen, R.G. and Simon, A. 1996. Geomorphological river channel reconnaissance for river analysis, engineering and management *Transactions of the Institute of British Geographers* 21: 469–483.

Thornes, J.B. 1977. Hydraulic geometry and channel change. In: Gregory, K.J., ed., *River Channel Changes*, Chichester: John Wiley & Sons, pp. 91–100.

Thornes, J.B. and Ferguson, R.I. 1981. Geomorphology. In: Wrigley, N. and Bennett, R.J., eds., *Quantitative Geography: a British View*, London: Routledge and Kegan Paul, pp. 284–293.

Tosi, M. 2007. Root tensile strength relationships and their slope stability implications of three shrub species in the Northern Apennines (Italy). *Geomorphology* 87: 268–283.

Tritton, D.J. 1988. *Physical Fluid Dynamics*, 2nd edn., Oxford: Oxford Science Publications.

Tucker, G.E. and Slingerland, R. 1997. Drainage basin responses to climate change. *Water Resources Research* 33: 2031–2047.

Van de Wiel, M.J. and Coulthard, T.J. 2010. Self-organized criticality in river basins: challenging sedimentary records of environmental change. *Geology* 38: 87–90.

Van de Wiel, M.J. and Darby, S.E. 2004. Numerical modeling of bed topography and bank erosion along tree-lined meandering rivers. In: Bennett, S.J. and Simon, A., eds., *Riparian Vegetation and Fluvial Geomorphology*, Washington, DC: American Geophysical Union, pp. 267–282.

Van de Wiel, M.J. and Darby, S.E. 2007. A new model to analyse the impact of woody riparian vegetation on the geotechnical stability of riverbanks. *Earth Surface Processes and Landforms* 32: 2185–2198.

Van de Wiel, M.J., Coulthard, T.J., Macklin, M.G. and Lewin, J. 2007. Embedding reach-scale fluvial dynamics within the CAESAR cellular automaton landscape evolution model. *Geomorphology* 90: 283–301.

Vasquez, J.A., Millar, R.G. and Steffler, P.M. 2011. Vertically-averaged and moment model for meandering river morphology. *Canadian Journal of Civil Engineering* 38: 921–931.

Vietz, G.J., Stewardson, M.J. and Finlayson, B.L. 2006. Flows that form: the hydromorphology of concave-bank bench formation in the Ovens River, Australia. In: Rowan, J.S., Duck, R.W. and Werrity, A., eds., *Sediment Dynamics and the Hydromorphology of Fluvial Systems*, IAHS Publication No. 306, Wallingford: International Association of Hydrological Sciences, pp. 267–276.

Viney, N.R. and Sivapalan, M. 1999. A conceptual model of sediment transport: application to the Avon River Basin in Western Australia. *Hydrological Processes* 13: 727–743.

Visconti, F., Camporeale, C. and Ridolfi, L. 2010. Role of discharge variability on pseudomeandering channel morphodynamics: results from laboratory experiments. *Journal of Geophysical Research* 115: F04042.

Vreugdenhil, C.B. 1994. Discretization in space. In: Vreugdenhil, C.B., ed., *Numerical Methods for Shallow-water Flow*, Dordrecht: Kluwer, pp. 89–116.

Wallerstein, N., Thorne, C.R. and Doyle, M.W. 1997. Spatial distribution and impact of large woody debris in northern Mississippi. In: Wang, S.S.Y., Langendoen, E.J. and Shields, F.D., eds., *Management of Landscapes Disturbed by Channel Incision*, Oxford, MS: University of Mississippi, pp. 145–150.

Wang S., White, W.R. and Bettess, R. 1986. A rational approach to river regime. In: Wang, S.S.Y., Shen, H.W. and Ding, L.Z., eds., *Proceedings of the 3rd International Symposium on River Sedimentation*, Oxford, MS: University of Mississippi, pp. 167–176.

Warburton, J. and Davies, T.R.H. 1998. Use of hydraulic models in management of braided gravel-bed rivers. In: Klingeman, P.C., Beschta, R.L., Komar, P.D. and Bradley, J.B., eds., *Gravel-bed Rivers in the Environment*, Highlands Ranch, CO: Water Resources Publications, pp. 513–542.

Werner, B.T. 2003. Modeling landforms as self-organized, hierarchical dynamical systems. In: Wilcock, P.R. and Iverson, R.M., eds., *Prediction in Geomorphology*, Washington, DC: American Geophysical Union, pp. 151–165.

Wheaton, J.M., Darby, S.E. and Sear, D.A. 2008. The scope of uncertainties in river restoration. In: Darby, S.E. and Sear, D.A., eds., *River Restoration: Managing the Uncertainty in Restoring Physical Habitat*, Chichester: John Wiley & Sons, pp. 21–39.

Whipple, K.X. and Dunne, T. 1992. The influence of debris-flow rheology on fan morphology, Owens Valley, California. *Bulletin of the Geological Society of America* 104: 887–900.

White, W.R., Bettess, R. and Paris, E. 1982. Analytical approach to river regime. *Journal of the Hydraulics Division, American Society of Civil Engineers* 108: 1179–1193.

Wiele, S.M. 1992. *A Computational Investigation of Bank Erosion and Midchannel Bar Formation in Gravel-Bed Rivers*, PhD Thesis, University of Minnesota, Minneapolis, MN.

Wilcock, P.R., Schmidt, J.C., Wolman, M.G., Dietrich, C.R., Dominick, D., Doyle, M.W., Grant, G.E., Iverson, R.M., Montgomery, D.R., Pierson, T.C., Schilling, S.P. and Wilson, R.C. 2003. When models meet managers: examples from geomorphology. In: Wilcock, P.R. and Iverson, R.M., eds., *Prediction in Geomorphology*, Geophysical Monograph, Washington, DC: American Geophysical Union, pp. 27–40.

Willgoose, G., Bras, R.L. and Rodriguez-Iturbe, I. 1991. A coupled channel network growth and hillslope evolution model. 1. Theory. *Water Resources Research* 27: 1671–1684.

Wolski, P., Savenije, H.H.G., Murray-Hudson, M. and Gumbricht, T. 2006. Modelling of the flooding in the Okavango Delta, Botswana, using a hybrid reservoir-GIS model. *Journal of Hydrology* 331: 58–72.

Wu, W., Shields, D., Bennett, S.J. and Wang, S.S.Y. 2005. A depth-averaged two-dimensional model for flow, sediment transport and bed topography in curved channels with riparian vegetation. *Water Resources Research* 41: W03015.

Xu, D., Bai, Y., Ma, J. and Tan, Y. 2011. Numerical investigation of long-term planform dynamics and stability of river meandering on fluvial floodplains. *Geomorphology* 132: 195–207.

Yang, C.T. 1971. Potential energy and stream morphology. *Water Resources Research* 7: 312–322.

Yang, C.T. and Song, C.C.S. 1979. Theory of minimum rate of energy dissipation. *Journal of the Hydraulics Division, American Society of Civil Engineers* 105: 769–784.

Yang, C.T., Song, C.C.S. and Woldenberg, M.J. 1981. Hydraulic geometry and minimum rate of energy dissipation. *Water Resources Research* 17: 1014–1018.

Yao, Z., Ta, W., Jia, X. and Xiao, J. 2011. Bank erosion and accretion along the Ningxia–Inner Mongolia reaches of the Yellow River from 1958 to 2008. *Geomorphology* 127: 99–106.

Yu, H.S., Salgado, R., Sloan, S.W. and Kim, J.M. 1998. Limit analysis versus limit equilibrium for slope stability. *Journal of Geotechnical and Geoenvironmental Engineering* 124: 1–11.

Zolezzi, G. and Seminara, G. 2001. Downstream and upstream influence in river meandering. Part 1. General theory and application to overdeepening. *Journal of Fluid Mechanics* 438: 183–211.

CHAPTER 18

Modelling flow, sediment transport and morphodynamics in rivers

Jonathan M. Nelson[1], Richard R. McDonald[1], Yasuyuki Shimizu[2], Ichiro Kimura[2], Mohamed Nabi[2] and Kazutake Asahi[3]

[1]US Geological Survey, Geomorphology and Sediment Transport Laboratory, Golden, CO, USA
[2]Hokkaido University, Sapporo, Japan
[3]RiverLink Corporation, Tokyo, Japan

18.1 Introduction

Overview

Predicting the response of natural or man-made channels to imposed supplies of water and sediment is one of the difficult practical problems commonly addressed by fluvial geomorphologists. This problem typically arises in three situations. In the first situation, geomorphologists are attempting to understand why a channel or class of channels has a certain general form; in a sense, this is the central goal of fluvial geomorphology. In the second situation, geomorphologists are trying to understand and explain how and why a specific channel will evolve or has evolved in response to altered or unusual sediment and water supplies to that channel. For example, this would include explaining the short-term response of a channel to an unusually large flood or predicting the response of a channel to long-term changes in flow or sediment supply due to various human activities such as damming or diversions. Finally, geomorphologists may be called upon to design or assess the design of proposed man-made channels that must carry a certain range of flows and sediment loads in a stable or at least quasi-stable manner. In each of these three situations, the problem is really the same: geomorphologists must understand and predict the interaction of the flow field in the channel, the sediment movement in the channel and the geometry of the channel bed and banks. In general, the flow field, the movement of sediment making up the bed and the morphology of the bed are intricately linked; the flow moves the sediment, the bed is altered by erosion and deposition of sediment and the shape of the bed is critically important for predicting the flow. This complex linkage is precisely what makes understanding channel form and process such a difficult and interesting challenge.

Until about the mid-1960s, channel form and response were evaluated primarily through qualitative understanding of a process coupled with detailed empirical observation. These approaches gave rise to several powerful tools that are still in use now, including regime theory and hydraulic geometry relationships. These tools provided geomorphologists with predictive methodologies for channel form and response. However, as the understanding of processes in channels has increased, so too has the detail of the questions being asked with regard to channel morphology and response to disturbance. Over the last few decades, the need for more precise predictive tools has led researchers in both geomorphology and engineering to formulate quantitative models of the coupled flow–sediment–bed system. These approaches are based on the capability to predict the flow field accurately, so their evolution in accuracy and detail over the last three decades has largely been determined by developments in computational methods for flow prediction. The techniques, which both complement and extend more classical techniques in fluvial geomorphology, offer powerful tools to geomorphologists trying to understand or predict stable channel forms and channel adjustments to altered flow and sediment supply.

In this chapter, a brief overview of techniques for predicting flow, sediment transport and bed evolution is presented, emphasizing the physical processes that are captured by various approaches. The goal of the chapter is not to provide recipes for constructing such models, although several components are discussed in detail and the industrious reader should find enough detail here and in the references to construct such a model. Rather, this material should be used as a guide in understanding these approaches and in selecting appropriate models for specific problems. All the models and model results presented in this chapter are available in the iRIC (International River Interface Cooperative, www.i-ric.org) public domain software, so readers have access to both the models and the tools used for generating imagery of model results. This is a rapidly developing field in geomorphology and engineering, so detailed discussion of specific models and algorithms is avoided for the most part, with the knowledge that most of these approaches are evolving over time and statements made herein about specific models may soon be outdated. On the other hand, the difference in processes captured by various approaches is emphasized, so

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

the reader may be able to judge which models or algorithms should be used for applications, both now and in the future, as more flexible models become available.

The coupled model concept

The key to the development of computational techniques for flow, sediment transport and bed evolution is the observation that, in most situations of practical interest, the time-scales associated with the flow are much shorter than those of bed and bank evolution. Hence, even when the bed is evolving, it does so slowly enough that the flow can be computed as if the bed and banks were not changing in time. This allows partial decoupling of the flow computation from the sediment motion and bed evolution. With this decoupling, it is possible to compute the flow field first, without simultaneously solving for the sediment-transport field and the bed morphology. One can compute the flow based on the input discharge (which may vary in time), use the flow solution to compute the sediment-transport patterns and evaluate those sediment-transport patterns to deduce local rates of erosion and deposition on the channel bed and banks. Given these rates and a specified time step, one can predict the topographic evolution some short time into the future. Provided that this time step is small enough, it is possible then to recompute the flow and continue to iterate on the flow, the sediment-transport field and the bed evolution, predicting the changes in each as a function of time. If the bed is not perfectly stable, but evolves in time, the flow patterns will change as time progresses even in the absence of discharge variations; they will change in response to the change in the channel morphology. Note that this intuitive methodology is the same across a range of actual modelling techniques from the simplest one-dimensional model to complex three-dimensional turbulence-resolving models; each exploits the separation in time-scales between the flow and the bed evolution to allow iterative, rather than simultaneous, solution of the governing relations. Therefore, although the general problem requires the simultaneous solution of the flow field, the sediment-transport field and the channel geometry, almost all practical problems can be solved with the much simpler iterative procedure.

Although the details of the methodology and specific applications have yet to be discussed here, the potential utility of the coupled model concept in geomorphology should be clear. The method allows one to examine the stability of a channel over time using hypothetical or real initial geometry, which is key to understanding both stable channel forms and the adjustment of channels to anthropogenic or natural changes in flow and sediment supply. The accuracy with which one can carry out these predictions depends critically on the choice of the various components of the coupled flow–sediment–bed modelling and on knowing what physics must be incorporated in the models to address certain classes of problems. With this in mind, this chapter deals with the particulars of such models with examples and hopefully will help readers to delineate the applicability and potential accuracy of various treatments.

18.2 Flow conservation laws

Conservation of mass and momentum

The conservation equations governing fluid and sediment motion are the fundamental building blocks of all coupled flow–sediment-transport–bed-evolution models, but various models use versions of the full equations that are reduced by neglecting certain terms or, more commonly, by integrating over one or more dimensions to develop averaged equations. The most important aspects to note in going through this exercise are the approximations that are required in order to develop certain methods; these will be explicitly noted in the text, as will the physical meaning of the approximations. The first approximation to be used here is that, throughout, the flow will be assumed to be incompressible. This is a good assumption provided that the flow velocities are much lower than the speed of sound, a condition that is well satisfied in channel flows. Using this assumption, conservation of mass and momentum for the flow is represented by the following equations (e.g. Tennekes and Lumley 1972):

$$\nabla \cdot \vec{u} = 0 \tag{18.1}$$

$$\frac{\partial \vec{u}}{\partial t} + \vec{u} \cdot \nabla \vec{u} = -\frac{1}{\rho} \nabla P + \vec{g} + \nu \nabla^2 \vec{u} \tag{18.2}$$

where

- $\vec{u}$ is the vector velocity, g is the gravitational constant,
- ρ is the fluid density,
- P is pressure and
- ν is the fluid kinematic viscosity.

These equations describe fluid motion in general; the only assumption made in deriving them is that the fluid is incompressible. In general, solving these equations in this full form in natural flows is difficult and impractical. Usually, the equations that are actually used to compute flow solutions are reduced forms of the above equations developed by temporal or spatial averaging or through scaling the equations to discover which terms are most important and retaining only those terms in the numerical solution.

The primary reason why these equations are difficult to solve for most natural flows is turbulence. With the exception of flows characterized by appropriate combinations of low velocity, small scale and/or high fluid viscosity [characterized by the Reynolds number; see Tennekes and Lumley (1972), pp. 1–26], flows are unstable to perturbations and are characterized by three-dimensional variability across a wide range of time and length scales. For example, even if one creates a simple channel flow with a smooth bottom, rectilinear channel shape and steady discharge, the velocity at any point in the flow will vary in time for typical length and time-scales due to turbulent eddies.

In addition to adding substantially to the complexity of the flow, these variations give rise to important momentum fluxes, changing even the time-averaged character of the flow significantly. To avoid the necessity of computing the variations in flow associated with turbulence, by far the majority of computational models used for natural flows use the so-called Reynolds equations. These equations are developed by splitting the vector velocity into a time-mean part (or an ensemble-averaged part) and a time-varying part (or the variation about the ensemble average). For a detailed description of this procedure and the reasoning behind it, the reader is referred to Tennekes and Lumley (1972, pp. 28–33) or any other beginning text on turbulence. In a Cartesian coordinate system with z positive upwards, the Reynolds momentum equations for the x, y and z directions are given by

$$\frac{\partial \overline{u}}{\partial t} + \overline{u}\frac{\partial \overline{u}}{\partial x} + \overline{v}\frac{\partial \overline{u}}{\partial y} + \overline{w}\frac{\partial \overline{u}}{\partial z} = -\frac{1}{\rho}\frac{\partial P}{\partial x} + \nu\nabla^2\overline{u} - \frac{\partial \overline{u'^2}}{\partial x} - \frac{\partial \overline{u'v'}}{\partial y} - \frac{\partial \overline{u'w'}}{\partial z} \tag{18.3}$$

$$\frac{\partial \overline{v}}{\partial t} + \overline{u}\frac{\partial \overline{v}}{\partial x} + \overline{v}\frac{\partial \overline{v}}{\partial y} + \overline{w}\frac{\partial \overline{v}}{\partial z} = -\frac{1}{\rho}\frac{\partial P}{\partial y} + \nu\nabla^2\overline{v} - \frac{\partial \overline{u'v'}}{\partial x} - \frac{\partial \overline{v'^2}}{\partial y} - \frac{\partial \overline{v'w'}}{\partial z} \tag{18.4}$$

$$\frac{\partial \overline{w}}{\partial t} + \overline{u}\frac{\partial \overline{w}}{\partial x} + \overline{v}\frac{\partial \overline{w}}{\partial y} + \overline{w}\frac{\partial \overline{w}}{\partial z} = -\frac{1}{\rho}\frac{\partial P}{\partial z} - g + \nu\nabla^2\overline{w} - \frac{\partial \overline{u'w'}}{\partial x} - \frac{\partial \overline{v'w'}}{\partial y} - \frac{\partial \overline{w'^2}}{\partial z} \tag{18.5}$$

where u, v and w are the velocity components in the x, y and z directions, respectively, overbars represent time (or ensemble) averages and primes represent deviations from that average (e.g. $u = \overline{u} + u'$). Strictly, time averaging would cause the first term in each momentum equation to be identically zero, but in practice, the time required to compute the average of a turbulent quantity is often less than the time-scale associated with externally imposed unsteadiness. For example, in a channel flow with slowly varying discharge, it may be possible to construct a time average over the turbulence using an averaging time much smaller than the time over which discharge variations occur. For ensemble averages, where one averages over many realizations of the same flow, the inclusion of the unsteady term in the equations is not problematic. For example, if one makes measurements of velocity in a turbulent wave boundary layer, it is possible to average over many waves to determine the ensemble-averaged behaviour of the flow; the departure from that average over a specific wave or time series of waves yields the turbulent variability. The last three terms on the right-hand side of the above equations arise as a result of the momentum fluxes due to turbulent fluctuations. These terms are very important for transferring momentum within the flow, especially near boundaries or anywhere strong shears occur in the flow.

Applying the same averaging procedure to the conservation of mass equation yields

$$\frac{\partial \overline{u}}{\partial x} + \frac{\partial \overline{v}}{\partial y} + \frac{\partial \overline{w}}{\partial z} = 0 \tag{18.6}$$

The original four equations expressing conservation of mass and momentum had four unknowns: the three components of velocity and the pressure. The number of unknowns matched the number of equations, so this was a well-posed problem. However, the four Reynolds-averaged mass and momentum equations yield more than four unknowns because of the appearance of the momentum fluxes associated with the turbulent fluctuations. This is the so-called closure problem of turbulence.

Reynolds stresses and turbulence closures

The quantities involving time or ensemble averages with products of time-varying quantities shown in eqns. 18.3–18.5 are referred to as Reynolds stresses. Although they are called stresses, it is important to remember that these terms arise due to advective transport of momentum. However, because they appear in the Reynolds-averaged momentum equations in a manner analogous to viscous stresses, they are referred to as stresses and are often parameterized in terms of the mean flow using concepts developed for viscous stresses. Rewriting eqns. 18.3–18.5 in terms of the components of the Reynolds stress tensor yields the following:

$$\frac{\partial \overline{u}}{\partial t} + \overline{u}\frac{\partial \overline{u}}{\partial x} + \overline{v}\frac{\partial \overline{u}}{\partial y} + \overline{w}\frac{\partial \overline{u}}{\partial z} = -\frac{1}{\rho}\frac{\partial P}{\partial x} + \nu\nabla^2\overline{u} + \frac{\partial \tau_{xx}}{\partial x} + \frac{\partial \tau_{yx}}{\partial y} + \frac{\partial \tau_{zx}}{\partial z} \tag{18.7}$$

$$\frac{\partial \overline{v}}{\partial t} + \overline{u}\frac{\partial \overline{v}}{\partial x} + \overline{v}\frac{\partial \overline{v}}{\partial y} + \overline{w}\frac{\partial \overline{v}}{\partial z} = -\frac{1}{\rho}\frac{\partial P}{\partial y} + \nu\nabla^2\overline{v} + \frac{\partial \tau_{xy}}{\partial x} + \frac{\partial \tau_{yy}}{\partial y} + \frac{\partial \tau_{zy}}{\partial z} \tag{18.8}$$

$$\frac{\partial \overline{w}}{\partial t} + \overline{u}\frac{\partial \overline{w}}{\partial x} + \overline{v}\frac{\partial \overline{w}}{\partial y} + \overline{w}\frac{\partial \overline{w}}{\partial z} = -\frac{1}{\rho}\frac{\partial P}{\partial z} - g + \nu\nabla^2\overline{w} + \frac{\partial \tau_{xz}}{\partial x} + \frac{\partial \tau_{yz}}{\partial y} + \frac{\partial \tau_{zz}}{\partial z} \tag{18.9}$$

where the Reynolds stresses are defined as follows:

$$\begin{aligned}
\tau_{xx} &= -\rho\overline{u'^2} \\
\tau_{yy} &= -\rho\overline{v'^2} \\
\tau_{zz} &= -\rho\overline{w'^2} \\
\tau_{xz} &= \tau_{zx} = -\rho\overline{u'w'} \\
\tau_{xy} &= \tau_{yx} = -\rho\overline{u'v'} \\
\tau_{yz} &= \tau_{zy} = -\rho\overline{v'w'}
\end{aligned} \tag{18.10}$$

Generally, the Reynolds stresses are much greater than viscous stresses in natural channel flows and the viscous stresses are neglected in the momentum equations. Hence the terms in the above equations involving ν, the kinematic viscosity, are negligibly small and are omitted from the equations.

In order to solve the above equations, one must either rewrite the Reynolds stresses in terms of the mean flow quantities or provide some other manner by which these terms may be evaluated using additional relations. The most common method in simulating natural flows is to relate the Reynolds stresses to the mean flow quantities by analogy with the relation between viscous stress and the rate of strain tensor. This leads to the concept of eddy viscosity, which assumes a proportionality between the Reynolds stresses and the components of the rate of strain. Although there is good justification for this kind of approach in situations where the flow is dominated by one length and velocity scale, as in a simple boundary layer, the concept is generally only a crude approximation for real, complex flows in nature. Nevertheless, many approaches are based on this concept and there are a number of ways of estimating the spatial structure and values for eddy viscosity using simple dimensional arguments or more complex reasoning. For example, some models use the eddy viscosity concept, but evaluate the local eddy viscosity using advection–diffusion equations for the turbulent kinetic energy and the length scale of the turbulence; this allows the treatment of situations where the local flow parameters are not accurate predictors of local turbulence structure. There are also a variety of closure approaches that are not predicated on the existence of an eddy viscosity. For example, it is possible to manipulate the momentum equations to develop expressions for each of the Reynolds stresses. However, these introduce more unknowns that must in turn be parameterized or estimated. A more complete discussion of turbulence closure techniques is beyond the scope of this chapter, but the reader is referred to the review by Rodi (1993) for an excellent discussion.

If the existence of a scalar, isotropic eddy viscosity, K, is assumed, the Reynolds stress terms in eqns. 18.7–18.9 may be replaced by the following relations:

$$\begin{aligned}
\tau_{xx} &\cong 2\rho K \frac{\partial \overline{u}}{\partial x} \\
\tau_{yy} &\cong 2\rho K \frac{\partial \overline{v}}{\partial y} \\
\tau_{zz} &\cong 2\rho K \frac{\partial \overline{w}}{\partial z} \\
\tau_{xz} = \tau_{zx} &\cong \rho K \left(\frac{\partial \overline{u}}{\partial z} + \frac{\partial \overline{w}}{\partial x} \right) \\
\tau_{xy} = \tau_{yx} &\cong \rho K \left(\frac{\partial \overline{u}}{\partial y} + \frac{\partial \overline{v}}{\partial x} \right) \\
\tau_{yz} = \tau_{zy} &\cong \rho K \left(\frac{\partial \overline{v}}{\partial z} + \frac{\partial \overline{w}}{\partial y} \right)
\end{aligned} \tag{18.11}$$

Substituting the above relations, eqns. 18.6–18.9 once again become a closed set of equations, with unknowns consisting of the Reynolds-averaged velocities and pressure. However, in order to solve these equations, an eddy viscosity still needs to be determined. As noted above, there are many ways to do this, but one of the most common is based on extending the well-posed relations for simple, steady, uniform boundary layers to more complex flows in channels. This extension is based on the observation that flows in unstratified channels are dominantly boundary layer-like in character. In simple boundary layers, the local turbulence is well described by the local boundary shear stress and distance from the boundary. Indeed, this result stems directly from simple dimensional analysis for steady, horizontally uniform flows (e.g. Tennekes and Lumley 1972). This result is complicated only slightly when one considers the effect of finite depth. The shear velocity is defined in terms of the local boundary shear stress and the fluid density as follows:

$$u_* = \left[\frac{(\tau_{zx})_{\mathrm{B}}}{\rho} \right]^{\frac{1}{2}} \tag{18.12}$$

where B denotes evaluation at the bed. Dimensional analysis yields the result that the eddy viscosity, K, can be written in the following form:

$$K = k u_* h \kappa(\xi) \tag{18.13}$$

where k is an empirical constant of proportionality called von Karman's constant (≈ 0.408; see Long *et al.* 1993) and $\kappa(\xi)$ is a shape function giving the vertical distribution of K between the bed and the water surface, using $\xi = z/h$, where h is the local flow depth and z is distance from the boundary. For the choice of a parabolic distribution of eddy viscosity, as given by

$$\kappa(\xi) = \xi(1 - \xi) \tag{18.14}$$

the velocity profile in the boundary layer will be logarithmic, as follows:

$$\overline{u} = \frac{u_*}{k} \ln\left(\frac{z}{z_0}\right) \tag{18.15}$$

where z_0, the so-called roughness length, is a constant of integration that depends on the boundary shear stress, the fluid viscosity and/or the size of the roughness elements on the bed [see Middleton and Southard (1984) or any text on wall-bounded shear flows for a discussion of roughness lengths]. In practice, experimental evidence suggests that eqn. 18.14 is not the best choice, although it may be fairly accurate close to the boundary. Although several other possibilities have been suggested in the literature, there is not much evidence to suggest that more complicated structure functions are verifiably better than simply using eqn. 18.14 from the bed up to one-fifth of the flow depth and using a constant value above that level, i.e.

$$\begin{aligned}
\kappa(\xi) &= \xi(1 - \xi) \qquad && \xi < 0.2 \\
\kappa(\xi) &= 0.16 && \xi \geq 0.2
\end{aligned} \tag{18.16}$$

This choice for κ yields a logarithmic velocity profile near the bed and a parabolic one well away from the bed and was first described by Rattray and Mitsuda (1974).

In applying models that use the simple eddy viscosity closure described above, it is absolutely critical to note that this form of the eddy viscosity is strictly correct only in a steady, uniform boundary layer. Although natural rivers and streams are predominantly boundary layer-like in nature and are commonly steady over time steps used in most models, they can be decidedly non-uniform, introducing free shear layers and wakes for which these eddy viscosity closures are inappropriate. One immediate shortcoming of the model above is that it predicts zero flux of momentum due to turbulence in regions where the boundary shear stress is zero. In a simple shear layer bounding a separation zone in a river, this suggests that, as the boundary shear stress must change sign somewhere in the region between upstream and downstream flow, there must be a surface across which no momentum is transferred by turbulence. This is wrong; if these effects are important, a different closure must be employed. Nevertheless, these simple closures perform adequately in a wide variety of natural flows. The most important point here is that, when using a closure of a certain type, one must keep in mind the potential errors in that closure and what physical processes are likely to be well treated and what processes are likely to be poorly treated.

Hydrostatic assumption

Up to this point, each of the three components of velocity has been treated equally and the terms in the momentum equations for u, v and w have been treated in the same manner. However, in many flows of interest, both vertical velocities and vertical accelerations are small and the vertical equation of motion (eqn. 18.5 or 18.9) can be accurately approximated by retaining only the pressure gradient and gravitational terms:

$$-\frac{1}{\rho}\frac{\partial P}{\partial z} - g = 0 \tag{18.17}$$

This assumption is referred to as the hydrostatic assumption, as it results in the pressure being distributed hydrostatically in the vertical, meaning that the pressure is equivalent to the overlying weight of fluid per unit area at any point. This simplification is a good one provided that vertical accelerations are small, meaning that bed slopes are relatively small along the direction of the flow. For flows with strong vertical acceleration produced by abrupt bed variations (as may be caused by bedrock or man-made structures), this assumption will be locally inaccurate, a point that will be revisited in a brief section below on fully three-dimensional models.

In situations where eqn. 18.17 is a suitable approximation for eqn. 18.9, the pressure gradients in the horizontal equations of motion can be written in terms of the water surface elevation, E, by integrating eqn. 18.17 in z and differentiating the result in each of the horizontal directions to obtain

$$-\frac{1}{\rho}\frac{\partial P}{\partial x} = -g\frac{\partial E}{\partial x} \tag{18.18}$$

$$-\frac{1}{\rho}\frac{\partial P}{\partial y} = -g\frac{\partial E}{\partial y} \tag{18.19}$$

These relations simplify the solution of the equations, because they reduce determining the pressure at each (x,y,z) location in the flow to determining only the water-surface elevation at each horizontal (x,y) location.

Coordinate systems

All of the above equations have been cast in a simple Cartesian coordinate system. In practice, flow solutions are computed in a wide variety of coordinate systems, including Cartesian, orthogonal curvilinear and general coordinate systems for finite difference solutions and a variety of structured and unstructured grids for finite element solutions. The primary advantage of general or unstructured grids is that they allow the coordinate system to be fitted precisely to the flow domain. The disadvantage is that they increase computational complexity considerably and, in cases where the bed and banks of the channel are evolving in time, the coordinate system must be recomputed at every time step, which is time consuming. In addition, at least some finite element solutions conserve mass only in a global sense; they typically are poor at enforcing mass conservation locally (Oliveira *et al.* 2000). This problem can be mitigated by careful construction of the flow grid, but it is difficult to avoid entirely, especially in channels with strong spatial accelerations produced by topography or channel curvature. Oliveira *et al.* (2000) found errors in local mass conservation of up to 85% after only 3 days of simulation applying standard finite element methods to the Tagus Estuary. In channel flows, errors of this magnitude result in solutions that are not good representations of the real flow and certainly could not be used to compute accurately the movement of sediment or other constituents within the flow.

Developing a variety of commonly used coordinate systems is not within the scope of this chapter, but it is worth mentioning one specific orthogonal curvilinear system that has been widely used in modelling river flows. This coordinate system is essentially a generalization of a cylindrical coordinate system where the curvature of the coordinate system is allowed to vary in the streamwise direction. This so-called 'channel-fitted' coordinate system has been used widely over the last 60 years or so, although most early applications involved only an incomplete set of equations. The system was formally derived and the full equations were published by Smith and McLean (1984). If the radius of curvature of the channel centreline is defined as R and s, n and z are defined as the streamwise, cross-stream and vertical coordinates, respectively, as shown in Fig. 18.1, the hydrostatic assumption is employed and the viscous stresses are assumed to be negligibly small and $N = n/R$, then the continuity

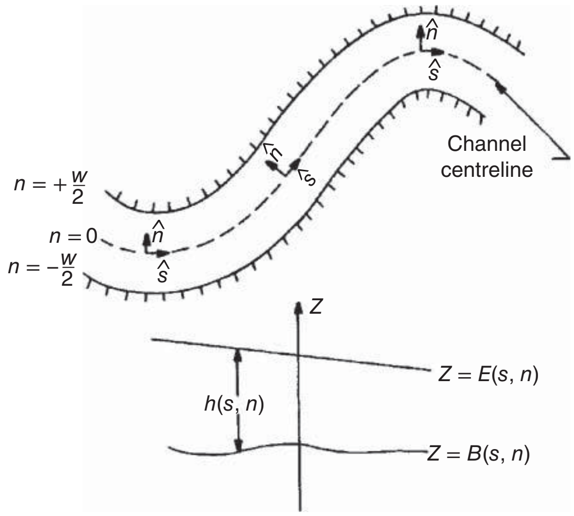


Figure 18.1 Schematic depiction of the curvilinear orthogonal coordinate system.

and momentum equations in this coordinate system are given by the following:

$$\frac{1}{1-N}\frac{\partial u}{\partial s} - \frac{v}{(1-N)R} + \frac{\partial v}{\partial n} + \frac{\partial w}{\partial z} = 0 \tag{18.20}$$

$$\frac{\partial u}{\partial t} + \frac{u}{(1-N)}\frac{\partial u}{\partial s} + v\frac{\partial u}{\partial n} + w\frac{\partial u}{\partial z} - \frac{uv}{(1-N)R} = \frac{-g}{1-N}\frac{\partial E}{\partial s} + \frac{1}{\rho}\left[\frac{1}{1-N}\frac{\partial \tau_{ss}}{\partial s} + \frac{\partial \tau_{ns}}{\partial n} + \frac{\partial \tau_{zs}}{\partial z} - \frac{2\tau_{ns}}{(1-N)\,R}\right] \tag{18.21}$$

$$\frac{\partial v}{\partial t} + \frac{u}{(1-N)}\frac{\partial v}{\partial s} + v\frac{\partial v}{\partial n} + w\frac{\partial v}{\partial z} + \frac{u^2}{(1-N)R} = \frac{-g}{1-N}\frac{\partial E}{\partial n} + \frac{1}{\rho}\left[\frac{1}{1-N}\frac{\partial \tau_{ns}}{\partial s} + \frac{\partial \tau_{nn}}{\partial n} + \frac{\partial \tau_{zn}}{\partial z} + \frac{\tau_{ss} - \tau_{nn}}{(1-N)\,R}\right] \tag{18.22}$$

$$-\frac{1}{\rho}\frac{\partial P}{\partial z} - g = 0 \tag{18.23}$$

If the existence of a scalar, isotropic eddy viscosity is assumed, we can rewrite eqn. 18.11 in the channel-fitted coordinate system, resulting in the following expressions for the six independent components of the deviatoric Reynolds stress tensor:

$$\begin{aligned}
\tau_{ss} &= 2\rho K\left[\frac{1}{1-N}\frac{\partial u}{\partial s} - \frac{v}{(1-N)\,R}\right] \\
\tau_{ns} &= \rho K\left[\frac{1}{1-N}\frac{\partial v}{\partial s} + \frac{u}{(1-N)\,R} + \frac{\partial u}{\partial n}\right] \\
\tau_{zs} &= \rho K\left(\frac{1}{1-N}\frac{\partial w}{\partial s} - \frac{\partial u}{\partial z}\right) \\
\tau_{nn} &= 2\rho K\left(\frac{\partial v}{\partial n}\right) \\
\tau_{zn} &= \rho K\left(\frac{\partial w}{\partial n} + \frac{\partial v}{\partial z}\right) \\
\tau_{zz} &= 2\rho K\left(\frac{\partial w}{\partial z}\right)
\end{aligned} \tag{18.24}$$

where the overbars denoting Reynolds averaging of the equations have been omitted for simplicity. If the radius of curvature of the channel centreline goes to infinity, meaning that the channel is straight, eqns. 18.20–18.24 revert back to the standard momentum equations with x and y oriented streamwise and cross-stream, respectively. However, if the channel is curved, the u and v velocity components in the s–n–z coordinate system still correspond to streamwise and cross-stream velocities, as the s-direction is always streamwise. Clearly, this would not be true if a Cartesian system were used; the orientation of the x and y components of velocity with respect to the channel would change with position. Thus, the channel-fitted coordinate system is in some sense the natural one, as it divides local velocity vectors into streamwise and cross-stream components. This system is also the one typically used in analysing field measurements in channels, because those measurements are frequently taken perpendicular to and parallel to sections that are themselves perpendicular to the channel centreline.

The first, and perhaps most confusing, step in applying the channel-fitted coordinate system is determining the channel centreline and the radius of curvature of that centreline. This is not a purely mathematical process; it requires some consideration of what one is trying to capture in the channel-fitted coordinate system. Provided that the numerics are correct and the full equations are used, the flow solution should be essentially independent of the coordinate system. Therefore, one could use a Cartesian coordinate system for a curved channel or even a curved coordinate system for a straight channel. However, if one chooses a coordinate system that follows the path of the channel, at least approximately, two advantages arise: first, the number of grid points required is minimized, and second, the convective accelerations associated with the curvature of the channel appear primarily in centripetal acceleration terms, rather than in differential terms in the governing equations. The latter consideration is the key to choosing the channel centreline for the coordinate system. Basically, one wants to find a centreline that captures the average curvature of the flow streamlines, which are approximately the same as the large-scale curvature of the banks. Because the flow 'averages' the effects of the local banks over a length scale comparable to width, one can digitize a centreline for the coordinate system (which need not correspond exactly to the channel centreline) in two ways: either one may digitize the centreline with points that are closer together than the channel width and then filter the resulting curve over distances of about a channel width, or one may simply choose a number of points, each about a channel width apart. In either case, the radius of curvature is easily found by noting that, if θ is the angle between the down valley direction and the local tangent to the centreline, the radius of curvature is given by

$$R = \left(\frac{\partial \theta}{\partial s}\right)^{-1} \tag{18.25}$$

When generating a channel-fitted coordinate system, the centreline defining the coordinate system should be drawn to approximate the average streamline curvature in the reach of interest as well as possible. It is not appropriate to take a precise channel centreline defined by a detailed (i.e. with spatial resolution much smaller than a channel width) survey of the banks, as the resulting detailed centreline may have local curvature values that are very poor approximations to the average streamline curvature.

Spatial averaging

In many cases, solution of the full momentum equations is not warranted either by the nature of the questions to be addressed in a given study or as a result of the kind and amount of data available. For example, applying a three-dimensional model to several hundred channel widths of a given river for a study of floodplain inundation when cross-sections of bathymetry are available only every 10 channel widths is not reasonable, because obtaining good results with a three-dimensional model would require more topographic data. Generally, more complete models that yield more precise results require much more input information in order to be applied relative to simpler models. In many cases, accurate results for a given purpose can be found using a simple model with relatively sparse topographic data. The two most common ways of developing simpler models are scaling analyses and spatial averaging. Scaling analysis refers to the concept of using the time and length scales of the flow to determine the most important terms in the governing equations and to develop simpler equations by retaining only these terms. This is a powerful tool for certain flows, but it generally results in a model that is specifically applicable to only a certain flow or class of flows. Spatial averaging is a method whereby one or more dimensions are removed from the model equations by integrating or averaging over those dimensions. For example, development of a one-dimensional flow model requires averaging the momentum equations over a channel cross-section, so that instead of solving for the velocity at every point in the channel, the model solves only for the cross-sectionally averaged velocity at each model cross-section. Note that although model simplicity is gained by spatial averaging, detail is lost.

The most common applications of spatial averaging result in one-dimensional models, two-dimensional models that treat the channel flow in planform (vertically averaged models) and two-dimensional models that treat the flow in the streamwise-vertical plane (cross-stream averaged models). Although treating each of these in any detail is beyond the scope of this introduction to modelling flow and sediment transport, a single example illustrates some of the issues that arise in developing spatially averaged equations. Using $\langle\rangle$ to represent vertical averaging, the vertical average of the u velocity component is defined as follows:

$$\langle u \rangle = \frac{1}{h}\int_B^E u\,\mathrm{d}z \tag{18.26}$$

Applying this same operator to eqns. 18.20–18.22, the following vertically averaged continuity and horizontal momentum equations arise in the channel-fitted coordinate system (again, note that the standard Cartesian relations are easily found from the following by letting R go to infinity):

$$\frac{1}{1-N}\frac{\partial}{\partial s}(\langle u\rangle h) - \frac{\langle v\rangle h}{(1-N)R} + \frac{\partial}{\partial n}(\langle v\rangle h) = 0 \tag{18.27}$$

$$\begin{aligned}\frac{1}{1-N}\frac{\partial}{\partial s}(\langle u^2\rangle h) + \frac{\partial}{\partial n}(\langle uv\rangle h) - \frac{2\langle uv\rangle h}{(1-N)R} &= -\frac{gh}{1-N}\frac{\partial E}{\partial s}\\ &+\frac{1}{\rho}\left[\frac{1}{1-N}\frac{\partial}{\partial s}\left(\left\langle \tau_{ss}\right\rangle h\right) + \frac{\partial}{\partial n}(\langle \tau_{ns}\rangle h) - \frac{2\langle \tau_{ns}\rangle h}{(1-N)R}\right]\\ &+\frac{1}{\rho}\left[\frac{1}{1-N}\left(\tau_{ss}\right)_B\frac{\partial B}{\partial s} + (\tau_{ns})_B\frac{\partial B}{\partial n} - (\tau_{zs})_B\right]\end{aligned} \tag{18.28}$$

$$\begin{aligned}\frac{1}{1-N}\frac{\partial}{\partial s}(\langle uv\rangle h) + \frac{\partial}{\partial n}(\langle v^2\rangle h) + \frac{(\langle u^2\rangle - \langle v^2\rangle)h}{(1-N)R} &= -\frac{gh}{1-N}\frac{\partial E}{\partial n}\\ &+\frac{1}{\rho}\left[\frac{1}{1-N}\frac{\partial}{\partial s}\left(\left\langle \tau_{ns}\right\rangle h\right) + \frac{\partial}{\partial n}(\langle \tau_{nn}\rangle h) - \frac{\langle \tau_{ss}-\tau_{nn}\rangle h}{(1-N)R}\right]\\ &+\frac{1}{\rho}\left[\frac{1}{1-N}\left(\tau_{ns}\right)_B\frac{\partial B}{\partial s} + (\tau_{nn})_B\frac{\partial B}{\partial n} - (\tau_{zn})_B\right]\end{aligned} \tag{18.29}$$

These equations, which have been used in variety of models for flow and bed evolution (Smith and McLean 1984; Nelson and Smith 1989a,1989b; Shimizu *et al.* 1991), introduce a new kind of closure problem that is analogous to the turbulence closure problem introduced by Reynolds averaging. Terms that arise due to vertical correlations such as $\langle uv\rangle$, $\langle u^2\rangle$ and $\langle v^2\rangle$ cannot be expressed in terms of simple vertically averaged variables such as $\langle u\rangle$ and $\langle v\rangle$ except where the velocities have no vertical structure whatsoever, so that $\langle uv\rangle = \langle u\rangle\langle v\rangle$ and $\langle u^2\rangle = \langle u\rangle^2$ and so forth. However, this is not generally true. For example, for a logarithmic velocity profile, the difference between $\langle u^2\rangle$ and $\langle u\rangle^2$ depends on the ratio of the roughness length to the flow depth and is typically on the order of 5–10%. In almost all vertically averaged models, the correlations are neglected and one assumes that the equalities that hold for the case of no vertical structure are accurate in cases with vertical structure. However, some important effects can be excluded when this assumption is used. For example, in long meander bends with weak topography, the term $\langle uv\rangle$ has been shown to be at least partially responsible for the movement of the high-velocity region of the flow from the inner bank at the upstream part of the bend to the outer bank at the downstream part of the bend (Shimizu *et al.* 1991). This is because helical cross-stream flow moves high-velocity fluid outwards near the surface of the flow and low-velocity fluid inwards near the bed, resulting in a net momentum flux towards the outer bank. This effect is overwhelmed by topographic steering of the flow in shorter bends with point bars, but it is potentially an important effect in some natural flows. Even though this effect is dependent on vertical structure, it can be treated to some extent in vertically averaged models using dispersion coefficients. Similarly, when spatial averaging is carried out, spatial correlations between variables that appear as a result of the averaging process can generally be treated at least to some approximate extent.

Dispersion coefficients

A general definition of a dispersion or correlation coefficient between two variables is given by the following:

$$\alpha_{ab} = \frac{\langle ab \rangle}{\langle a \rangle \langle b \rangle} \tag{18.30}$$

where $\langle\rangle$ may represent vertical averaging or some other spatial average (e.g. cross-sectional). Using this definition, eqns. 18.28 and 18.29 may be rewritten in terms of only $\langle u \rangle$ and $\langle v \rangle$ along with the dispersion coefficients α_{uu}, α_{vv} and α_{uv}. The values of these coefficients may be set theoretically or empirically. In either case, the coefficients allow at least approximate treatment of momentum fluxes that would otherwise be neglected. Another way to treat the correlation terms in averaged equations is to separate each variable into an averaged part and a deviation from that average, in parallel with the development of the Reynolds momentum equations. For example, if we use primes to denote departures from the vertical average, such as $u(z) = \langle u \rangle + u'(z)$, we can rewrite eqns. 18.28 and 18.29 as follows:

$$\begin{aligned} &\frac{1}{1-N}\frac{\partial}{\partial s}(\langle u \rangle^2 h) + \frac{\partial}{\partial n}(\langle u \rangle \langle v \rangle h) - \frac{2\langle u \rangle \langle v \rangle h}{(1-N)R} \\ &\quad + F' = -\frac{gh}{1-N}\frac{\partial E}{\partial s} \\ &+\frac{1}{\rho}\left[\frac{1}{1-N}\frac{\partial}{\partial s}\left(\left\langle \tau_{ss}\right\rangle h\right) + \frac{\partial}{\partial n}(\langle \tau_{ns} \rangle h) - \frac{2\langle \tau_{ns} \rangle h}{(1-N)R}\right] \\ &+\frac{1}{\rho}\left[\frac{1}{1-N}\left(\tau_{ss}\right)_B \frac{\partial B}{\partial s} + (\tau_{ns})_B \frac{\partial B}{\partial n} - (\tau_{zs})_B\right] \end{aligned} \tag{18.31}$$

$$\begin{aligned} &\frac{1}{1-N}\frac{\partial}{\partial s}(\langle u \rangle \langle v \rangle h) + \frac{\partial}{\partial n}(\langle v \rangle^2 h) + \frac{(\langle u \rangle^2 - \langle v \rangle^2)h}{(1-N)R} \\ &\quad + G' = -\frac{gh}{1-N}\frac{\partial E}{\partial n} \\ &+\frac{1}{\rho}\left[\frac{1}{1-N}\frac{\partial}{\partial s}\left(\left\langle \tau_{ns}\right\rangle h\right) + \frac{\partial}{\partial n}(\langle \tau_{nn} \rangle h) - \frac{\langle \tau_{ss} - \tau_{nn} \rangle h}{(1-N)R}\right] \\ &+\frac{1}{\rho}\left[\frac{1}{1-N}\left(\tau_{ns}\right)_B \frac{\partial B}{\partial s} + (\tau_{nn})_B \frac{\partial B}{\partial n} - (\tau_{zn})_B\right] \end{aligned} \tag{18.32}$$

where the new terms are defined by

$$F' = \frac{1}{1-N}\frac{\partial}{\partial s}(\langle u'^2 \rangle h) + \frac{\partial}{\partial n}(\langle u'v' \rangle h) - \frac{2(\langle u'v' \rangle)h}{(1-N)R} \tag{18.33}$$

and

$$G' = \frac{1}{1-N}\frac{\partial}{\partial s}(\langle u'v' \rangle h) + \frac{\partial}{\partial n}(\langle v'^2 \rangle h) + \frac{(\langle u'^2 \rangle - \langle v'^2 \rangle)h}{(1-N)R} \tag{18.34}$$

In cases where simple structure functions can be supplied for u and v based on measurements or theoretical arguments, these 'extra' terms arising from correlations can be evaluated approximately. If these terms are set to zero, it is important to have an understanding of what kinds of processes are being neglected in the formulation. Situations where spatial correlations are important can often be treated without solving the full equations.

Bed stress closure

Whenever the equations of motion are averaged in the direction perpendicular to a boundary, closures for stress terms at that boundary must be supplied. In the vertically averaged equations used as an example, the boundary shear stress terms that arise in the horizontal momentum equations must be expressed in terms of $\langle u \rangle$ and $\langle v \rangle$. There are many ways to do this, including using Manning's or Chezy's closure, as discussed below, but the most common in multidimensional models is to use a drag coefficient (C_d) closure:

$$\tau_B = \rho C_d (u^2 + v^2) \tag{18.35}$$

Splitting this into component parts yields

$$(\tau_{zs})_B = \rho C_d \sqrt{\langle u \rangle^2 + \langle v \rangle^2}\langle u \rangle \tag{18.36}$$

and

$$(\tau_{zn})_B = \rho C_d \sqrt{\langle u \rangle^2 + \langle v \rangle^2}\langle v \rangle \tag{18.37}$$

There are many other choices of bottom stress closure, but most can be directly related to this one. For example, if the flow is assumed to have a vertical structure:

$$u = u_* f(z, z_0) \tag{18.38}$$

For this case, the drag coefficient can be shown to be a function only of flow depth and z_0:

$$C_d = \left[\frac{1}{h}\int_{z_0}^{h} f\left(z, z_0\right) \mathrm{d}z\right]^{-2} \tag{18.39}$$

Closures for lateral shear stresses at banks can be handled in a similar manner. Using this closure or others that are similar, the vertically averaged horizontal momentum equations and the continuity equation can be written entirely in terms of the vertically averaged u and v velocity components and the water-surface elevation (if the flow is assumed to be hydrostatic). This is a well-posed system of equations and unknowns, so a solution is straightforward. Although these assumptions are often not explicitly stated, any model developed from spatial averaging of the full equations requires specification of dispersion coefficients and closures for stresses at boundaries.

18.3 Sediment-transport relations

In order to determine the rates of transport of sediment travelling as bedload or in suspension, information from the flow model is typically used as input to a variety of empirical, semiempirical or theoretical relations for predicting sediment flux. Computations of local fluxes can be used with the equation for conservation of sediment mass to predict local erosion and deposition. However, relatively small errors in local fluxes can make a significant difference in the local rates of erosion and deposition and errors in methods for computing sediment fluxes are often large. Choosing a method that can be calibrated with measured data, or that was developed in situations with similar grain sizes and flow characteristics, is the best way to build confidence in predictions. As in most complex problems in physical science, progress is almost always made in the interplay between careful field measurement and modelling efforts.

Bedload transport

Bedload transport refers to grain motion near the bed consisting of rolling and hopping grains; these grains typically are moving with horizontal velocities less than the speed of the flow through most of their trajectory. Although there have been a few notable attempts to develop purely theoretical relations for bedload sediment entrainment and motion, even these models rely heavily on empirical data and most predictions of bedload flux are made using empirical equations. As a result, it is especially important to understand how a given relation was developed and calibrated when choosing a method for computational prediction. Papers by Gomez (1991) and Recking (2010) provide a good overview of methods for predicting and measuring bedload transport and also point out some of the physical characteristics that make developing a general model difficult. Equations for predicting bedload flux as a function of properties of the flow (velocity, boundary shear stress, stream power, viscosity, fluid density, etc.) are usually dependent on grain size and density and may also depend on sorting or other properties of the bed itself. One of the simplest bedload equations was developed by Meyer-Peter and Müller (1948) and it will serve as an example of these equations for the purposes of this chapter. Defining non-dimensional transport and boundary shear stress as follows:

$$(q_b)_* = \frac{q_b}{\left[\left(\frac{\rho_s - \rho}{\rho}\right) g D^3\right]^{\frac{1}{2}}} \tag{18.40}$$

Where

q_b is the volumetric bedload flux per unit width,
d is the grain size,
g is the gravitational constant,
ρ_s is the sediment density,
ρ is the fluid density and

$$\tau_* = \frac{\tau_b}{[(\rho_s - \rho) g D]} \tag{18.41}$$

the Meyer-Peter and Müller (1948) equation can be written as

$$(q_b)_* = 8(\tau_* - 0.047)^{\frac{3}{2}} \tag{18.42}$$

Hence the Meyer-Peter and Müller (1948) bedload equation yields the bedload flux as a function of only the boundary shear stress, grain size and the particle and fluid densities. Many users apply the so-called modified Meyer-Peter and Müller equation, given by

$$(q_b)_* = 8[\tau_* - (\tau_*)_c]^{\frac{3}{2}} \tag{18.43}$$

where $(\tau_*)_c$ is the non-dimensional form of the Shields critical shear stress. The Shields critical shear stress is defined as that value of shear stress for which significant sediment motion begins to occur for a given grain size. The reader is referred to Middleton and Southard (1984) and Garcia (2008) for an in-depth review of this quantity and methods for determining the value of critical shear stress. Many other bedload equations also use this concept. Although critical shear stress was originally developed for the case of well-sorted beds that could be considered uniform in size, the concept has been generalized and extended to the case of mixed grain-size beds by several researchers (e.g. Wiberg and Smith 1987). Using a critical shear stress developed for beds of mixed sizes is the commonest way to deal with poorly sorted sediment beds in sediment-transport models, but it is important to note that this treatment does not correctly parameterize many of the details of mixed-grain transport. This is especially true if small-scale spatial sorting occurs or if the texture or structure of the bed evolves during flow events in other ways. Recent progress on more complete parameterization of mixed-grain transport appears likely to lead to better models. For further discussion on this and related topics, the reader is referred to Wilcock (1997, 2001) and a review by Parker (2008).

Suspended load transport

Suspended load is carried by the flow both near and well above the bed, depending on the grain size and the turbulence levels in the flow, as characterized by the Rouse number [see, for example, Middleton and Southard (1984)]. Sediment particles moving in suspension travel at approximately the horizontal speed of the flow. In many rivers and streams, suspended load, which is typically finer and faster moving than bedload, is a greater contributor to the overall sediment load of the channel than bedload. However, the bedload is often still very important for understanding the geomorphology of the channel, because permanent bed and bank features are often dominantly made up of the grain sizes carried as bedload. Furthermore, the quantity of suspended load may not be as tightly coupled to the hydraulics (flow characteristics) of the channel compared with bedload, because the amount of suspended material in transport may be governed primarily by the amount of fine material supplied to the channel. Thus, hysteresis in suspended load relations is much more common than in bedload relations. This characteristic can make suspended load more difficult to estimate, especially for the finest sizes in suspension.

In some cases, it is possible to calculate the flux of suspended load using an empirical total load equation, such as that proposed by Engelund and Hansen (1967). However, in most cases, models use some form of the advection–diffusion equation to treat suspended load transport. As in the case of momentum, turbulence produces advective transport of suspended sediment. Following Reynolds averaging and assuming a gradient-transport closure with a scalar isotropic eddy viscosity, the advection–diffusion equation for suspended sediment in vector form is given by

$$\frac{\partial c_s}{\partial t} + (\vec{u} - \vec{w}_s) \cdot \nabla c_s = \nabla \cdot K \nabla c_s \tag{18.44}$$

where c_s is the concentration of suspended material, $\vec{w}_s$ is the settling velocity (positive downwards) and K is the eddy diffusivity. For steady, uniform flow and an eddy diffusivity of the form given in eqn. 18.14, this equation can be solved directly to yield the Rouse profile [see Middleton and Southard (1984), p. 219], provided that an appropriate lower boundary condition for the sediment concentration is supplied. For more complex flows, the flow solution can be inserted along with the appropriate diffusivity and the equation can be solved numerically for the distribution of suspended sediment. For steady, uniform flows, the boundary condition at the bed is generally taken as a simple reference concentration [as a function of boundary shear stress, for example, see Garcia and Parker (1991)]. For more complex flows, the lower boundary condition is set by using a boundary condition on upward flux from the bed as a function of boundary shear stress. The form of the reference flux condition for non-uniform flows is derived directly from generalizing the reference concentration for uniform flows into an upward flux boundary condition. Thus, for example, in a situation where the boundary shear stress goes to zero at a point in a non-uniform flow, the upward flux off the bed is assumed to be zero and the actual concentration at the bed is set by the settling of grains already in suspension in the flow.

In situations with both high concentrations and high concentration gradients, corrections to the eddy diffusivity must be made due to the stratifying effect of the suspended sediment. The reader is referred to McLean (1992) for an in-depth discussion of stratification corrections.

Erosion equation

Once the flux of bedload and suspended load have been computed, determination of the local erosion or deposition on the bed is straightforward. Applying conservation of sediment mass, the rate of erosion or deposition on the bed is given by the so-called erosion equation:

$$\frac{\partial B}{\partial t} = -\frac{1}{c_b}\left[\nabla \cdot \vec{Q}_s + \frac{\partial}{\partial t}\int_B^E c_s \mathrm{d}z\right] \tag{18.45}$$

where $\vec{Q}_s$ is the local vector sediment flux and c_b is the concentration of sediment in the bed (typically about 0.65, i.e. unity minus the porosity).

Gravitational corrections to sediment fluxes

When sediment moves as bedload over a laterally sloping bed, the sediment will not move in the direction of the near-bed flow and bottom stress, but will be deflected somewhat downslope due to the action of gravity. The degree of deflection is roughly related to the ratio of drag forces on the particle and gravitational forces on the particle, with low values of that ratio corresponding to greater downslope deflection of the particle path. Because gravitational forces are proportional to particle volume, whereas drag forces are proportional to particle area, larger particles typically experience greater deflections than smaller particles. This explains, for example, why coarse grains are preferentially sorted down the sloping faces of point bars relative to finer particles. There are several published gravitational correction models and all are fairly similar. Nelson (1990) showed that the bedload gravitational corrections developed by Engelund (1974), Kikkawa *et al.* (1976), Hasegawa (1984) and Parker (1984) could all be written in the following form:

$$Q_n = Q_s\left[\frac{\tau_s}{\tau_n} + \Gamma f\left(\frac{\tau_c}{\tau_b}\right)\frac{\partial B}{\partial n}\right] \tag{18.46}$$

where τ_s and τ_n refer to the streamwise and cross-stream components of the boundary shear stress, Q_s and Q_n are the streamwise and cross-stream components of bedload sediment flux, Γ is a coefficient and f is a simple function of the ratio of critical to boundary shear stress. For details of the values of Γ and f, the reader is referred to Nelson (1990) or the original publications listed above. These corrections were all developed assuming that no correction needs to be made along the direction of the boundary shear stress, but this assumption is questionable and awaits more careful experimental examination. Nelson (1990) proposed a method of gravitational correction based on the creation of a gravitational pseudo-stress that is added in a vector sense to the boundary shear stress. This formulation also reduces to eqn. 18.46 for the case of small angles and cross-stream corrections only, but also treats corrections in an approximate manner for bed slopes oriented arbitrarily with respect to the boundary shear stress.

Gravitational corrections are extremely important in bed evolution models as they play a critical role in determining the lateral slopes of bars. Unless transport, erosion and deposition are completely dominated by suspended load, a correction for the influence of gravity is a necessity for accurate prediction of bar morphology.

18.4 Numerical methods

A full discussion of the various numerical methods used in computing flow, sediment transport and bed evolution would be difficult to cover even in a book, much less a chapter or a chapter section. Because the intent of this book is to provide an overview of tools in geomorphology, not tools in computational fluid mechanics, the subject of numerical techniques will be given short shrift here, although certain common algorithms will be referred to briefly in subsequent sections. Nevertheless, this is an important part of constructing coupled models for predicting channel behaviour and particular care must be taken in choosing algorithms. There are two primary issues, somewhat related, that require special attention in choosing algorithms: stability and numerical dispersion.

Stability, or more precisely the lack of it, is easy to observe in model results. Poorly designed algorithms for computing flow and/or bed evolution lead to unrealistic results that rapidly become more unrealistic as one iterates towards a steady solution or steps the model forward in time for unsteady solutions. Stability considerations for the flow computations alone are generally outlined by the author of the flow computation method. Stability considerations for coupled flow–sediment–bed models are altogether more subtle and depend on a number of considerations. First, the time step of bed evolution must be chosen such that bed evolution is slow relative to the time-scales associated with the flow field, as this is really the basic premise of the semi-coupled modelling approach. If large changes in the bed and/or bank geometry occur within a single flow time step, the solution is almost certain to be unstable. Second, the numerical techniques must be chosen such that artificial phase lags between flow and sediment parameters are not introduced. This may seem complicated but actually relies on basic common sense. Consider the following example: if a one-dimensional model is used on a low-Froude number flow through a simple channel constriction, the cross-sectionally averaged velocity (which is all one computes in a true one-dimensional channel model) will be maximum at the constriction. If that velocity is used to compute bedload sediment transport, it will also be a maximum at the constriction, assuming typical relations between velocity, bed stress and sediment flux. Because the flux is maximum at the constriction, the spatial gradient in sediment flux is zero at that point. Because the spatial gradient of the flux is directly related to erosion and deposition (eqn. 18.45), the constriction will neither expand nor contract further. However, noting that the flux must be less than the value at the constriction both upstream and downstream of it, a paradox arises. If the spatial gradient in the flux is computed at the constriction throat using the value at the throat and the one immediately upstream, erosion is predicted to occur at the constriction. If the value at the constriction and the value immediately downstream are used, deposition is predicted to occur at the constriction. Both results are wrong and will lead to runaway expansion or contraction of the constriction. This can be dealt with in a number of simple ways, but the example shows how phase lags introduced between the flow and sediment transport parameters can lead to instabilities in the bed that are not real. Numerical methods must be chosen to avoid artificial instability of the flow field as well as the coupled flow–bed–sediment system.

Excessive numerical dispersion is typically not as obvious to the user as a stability problem. One of the important physical elements of modelling flow and sediment transport is the treatment of the movement of mass and momentum due to true diffusion or to advective processes that can be treated as diffusion-like (notably the transfer of momentum and mass by turbulence). Although a detailed mathematical discussion of this topic is outside the scope of this chapter, one of the basic problems of treating continuous systems with discretized equations is that commonly some artificial transfer of mass and/or momentum can occur as a result of the discretization process. This is referred to as numerical dispersion or numerical viscosity. The magnitudes of these effects are strongly dependent on the numerical scheme chosen and the actual numerical grid. Ideally, one would like numerical dispersion to be vanishingly small relative to the real processes of dispersion that one is trying to treat in the numerical solution, thereby ensuring that the model results are consistent with real-world observations. Unfortunately, numerical dispersion has an added benefit for models that tend to be unstable in that it effectively increases the stability of the model solutions. Accordingly, it is not unusual to see model results where the values of diffusivities are an order of magnitude (or more) larger than real-world values, where the unrealistically high values are assigned strictly to provide model stability. These models produce artificially smooth distributions of velocity and stress and generally cannot provide accurate predictions of sediment flux or bed morphology. The hallmarks of this kind of approach for two- or three-dimensional models are separation eddies that are very short relative to real-world values, rapid spreading of shear layers in the streamwise direction and near-bank shears that are low relative to observations. Typically, models with very large values of numerical dispersion show insensitivity to the parameters of the model governing momentum exchange (e.g. drag coefficient, Manning's *n*, turbulent diffusivity). Models that use unrealistically high values of diffusivity often are unable to produce stable solutions when using realistic values of diffusivity.

Although the problems of stability issues and numerical dispersion are especially important in coupled models for flow, sediment transport and bed evolution, there are many other considerations to be made in developing numerical techniques for such approaches. Fortunately, there are many excellent texts on this subject; for specific examples of different numerical solution techniques, the reader is referred to the texts by Patankar (1980), Chaudhry (1993) and Abbott and Minns (1998). Furthermore, for well-written algorithmic elements that are useful in a variety of different approaches (e.g. tridiagonal solvers, matrix inverters, alternating direction implicit solvers, mesh generators, etc.), the reader is encouraged to explore *Numerical Recipes* (Press *et al.* 1986) and the algorithms in current libraries of standard applications (e.g. IMSL, Matlab, Mathematica).

18.5 One-dimensional models

As has already been pointed out, in many cases, solution of the full momentum equations is not warranted by either the nature of the questions to be addressed in a given study or by the kind and amount of data available. As noted above, applying a three-dimensional model to several hundred channel widths of a river for a study of floodplain inundation when cross-sections of bathymetry are available only every 10 channel widths is not reasonable. In this situation, a one-dimensional model is probably more appropriate. Development of a one-dimensional flow model requires averaging the momentum equations over a channel cross-section, so that instead of solving for the velocity at every point in the channel, the model solves only for the cross-sectionally averaged velocity, flow rate or discharge at each model cross-section. Recall that although model simplicity is gained by spatial averaging, detail is lost. Nevertheless, one-dimensional models are suitable for a wide range of important problems and they are simple to develop and use. Because the central topic of this chapter is using models in fluvial geomorphology, which specifically involves the prediction of sediment transport and channel form (for both of which one-dimensional approaches are poorly suited), one-dimensional approaches are only briefly mentioned here.

One-dimensional processes

One-dimensional models capture a relatively small fraction of the processes that are active in rivers and streams, but the key to their overall success and utility is that they can make predictions over long length and time-scales. Because these models predict only cross-sectionally averaged quantities, they cannot predict vertical or cross-stream flow structure. They handle the response of the flow to expansions and contractions in the channel fairly well, correctly predicting the streamwise free-surface response to these features. One of the most common uses of one-dimensional models is for predicting water surface levels for various hydrographs, and these techniques are still the most commonly used for predicting inundation levels during flood events. Because they treat flow expansion and contraction well, one-dimensional mobile-bed models are appropriate for determining cross-sectionally averaged scour or fill.

There are a variety of one-dimensional models that incorporate two-dimensional processes through empirical relations. Generally, these models are applicable for the situations for which they are calibrated and they can be useful when carefully applied, but extending them outside their immediate range of applicability is prone to error. Typically, models that attempt to treat two-dimensional processes (such as bar formation) introduce several additional coefficients or parameters and are often more complex than a simple two-dimensional approach. An example of this is the so-called stream tube method, where the flow in a channel is reduced to one-dimensional flow in a suite of stream tubes that span the channel. Coefficients or parameters accounting for momentum exchange between the tubes must be incorporated and it is questionable whether these models are any simpler than a more correct two-dimensional application, which the present authors would recommend.

One-dimensional models

A number of public domain one-dimensional flow models are available. In general, these are appropriate for simulating quasi-steady or fully unsteady vertically homogeneous flow in networks of interconnected one-dimensional channels such as rivers with tributaries, tidally influenced barge canals and delta distribution systems. For example, the US Army Corps of Engineers (1997) provides a popular program called HEC-RAS (see http://www.hec.usace.army.mil/) that provides the user with the ability to apply several external and internal boundary conditions, including flow and stage hydrographs, rating curves, gated and uncontrolled spillways, pumps and bridges and culverts. The US Geological Survey (USGS) has three models (see http://water.usgs.gov/software/) that can be applied in similar situations. These include BRANCH (Schaffranek 1987), FEQ (Franz and Melching 1997*)* and FOURPT (DeLong *et al.* 1997). BRANCH is generally used by the USGS for computing discharge at backwater-affected stream gauging stations. The public domain iRIC software package (www.i-ric.org) currently includes the one-dimensional model CERI-1d, which also incorporates a unique extension to ice coverage and breakup. Generally, these models are for treating flow but the extension to treating flow, sediment transport and bed evolution is straightforward and is incorporated in some of the above models. The reader can see a more comprehensive list of one-dimensional models with sediment transport in the reviews by Fan (1993) and Thomas and Chang (2008). However, owing to the limitation of one-dimensional models in predicting the evolution of bars and banks, they are generally of limited value for the prediction of morphodynamics, although they are frequently used for routing of sediment using measured rating curves or other data-driven techniques.

18.6 Two-dimensional models

In many cases where spatial detail and specific bar forms are not of interest, one-dimensional models may efficiently represent large-scale flow and sediment-transport processes. However, if specific questions about at-a-point flow, sediment transport and erosion and deposition must be answered, a two- or three-dimensional model is required. For example, if the questions to be addressed are related to the position and amplitude of bars within the channel reach of interest, generally a two-dimensional model is necessary, as a one-dimensional model cannot predict the local flow and transport structure that gives rise to bar evolution. Similarly, if the flow field of interest includes steering of the flow around islands or bars or if there is significant cross-stream variability in the flow, at least a two-dimensional model should be applied to predict the details of local sediment transport or changes in bed morphology. In some cases where one-dimensional models yield cross-sectionally averaged velocities that are incapable of entraining sediment, two-dimensional computations will show a high-velocity region of flow in the channel where sediment is in motion.

Two-dimensional processes

In going from a one- to a two-dimensional model, three critical improvements are gained. First, instead of predicting only the cross-sectionally averaged component of downstream velocity and bed stress, the model predicts the value of vertically averaged downstream velocity and bed stress at many points across the channel. This means that the model can explicitly treat situations with large cross-stream velocity gradients and flow separation, which is particularly important when computing sediment transport. Second, the model also predicts the cross-stream components of vertically averaged velocity and bed stress at each point in the computational grid. As already noted, this means that a two-dimensional model can handle steering of the flow around bars and islands. This capability is critically important for the prediction of the evolution and stability of bars in rivers, as the basic instability leading to these is often associated with the interaction of topographic steering of the flow and the sediment transport (Nelson and Smith 1989b). Finally, two-dimensional models allow the prediction of cross-stream structure in the water-surface elevation, whereas one-dimensional approaches do not. In many cases, superelevation of the water surface due to channel curvature or bathymetric variability results in cross-stream gradients in water-surface elevation that are much larger than downstream components, so if accurate local water-edge elevation is required, at least a two-dimensional model should be applied.

These three basic enhancements have several corollaries. Because it yields spatially localized quantities, a two-dimensional approach can be used to predict near-bank velocities, stresses and sediment evacuation rates, which may be critical for predicting bank erosion. In addition, because these approaches give detailed information in a planform sense, they are useful for evaluating fields other than simply sediment transport and bed evolution. For example, two-dimensional approaches are currently becoming the standard for habitat modelling. Habitat evaluation for many riparian species typically requires physical variables including vertically averaged velocity, depth, substrate and so forth. A one-dimensional model could only evaluate habitat on a cross-sectional basis, which is not sufficient in most cases, as streamwise variations are typically unimportant relative to lateral variations.

It is important to point out that the enhanced predictive capabilities of a two-dimensional approach do not come without a price. Typically, the input data (primarily topography) required for two-dimensional modelling application are substantially more detailed than those required for a one-dimensional model. Two-dimensional models are also more computationally intensive and require a great deal more field data for verification or testing. In situations where one-dimensional models are sufficient for answering the research question or where data of sufficient detail to warrant a two-dimensional application are not available, there may be no point in applying the two-dimensional approach.

Two-dimensional models

A wide variety of steady and unsteady two-dimensional flow models are available. In the interest of providing access for the reader to any models used or discussed in this chapter, the public-domain open-source models incorporated in the iRIC interface are discussed here. These models are described in tabular form in Table 18.1 later in the chapter. The iRIC interface, the listed models, tutorials, users' manuals and other information are freely offered on the iRIC website at www.i-ric.org. The available models include both finite difference and finite element solutions, with the advantages and disadvantages already discussed above in the section on coordinate systems.

Although most current two-dimensional models are fully unsteady, meaning that the unsteady terms are retained in the momentum equations, some two-dimensional mobile bed models (e.g. the two-dimensional version of FaSTMECH in the iRIC suite of models) can handle hydrographs by varying the discharge in time without including the unsteady term in the momentum equations (the flow is assumed to be 'quasi-steady'). This is a reasonable assumption only if the unsteady term in the momentum equation can be shown to be small relative to the other terms in the equation. Although quasi-steady models cannot be used to simulate situations with rapidly varying discharge (e.g. flash floods or dam breaks), they are computationally much less demanding than fully unsteady models.

Although there are many two-dimensional flow models freely available for use, only a few of these are coupled flow, sediment transport and bed evolution models; because the aim here is to show how models can be used in fluvial geomorphology, these 'morphodynamics' models are discussed here. The sediment-transport components of two-dimensional (vertically averaged) models are developed as described above, with sediment fluxes computed in both streamwise and cross-stream directions. A gravitational correction is typically incorporated. Developing a vertically averaged solution for suspended sediment flux is difficult except for the simplest case of vanishingly small Rouse numbers. There are essentially three common techniques for treating this issue. First, some two-dimensional models that treat suspended sediment advection–diffusion do so by assigning vertical structure functions for velocity and solving the three-dimensional advection–diffusion equation for sediment concentration. Second, some models assign vertical structure functions for both velocity and sediment concentration and then solve a vertically averaged version of eqn. 18.44 using the vertical structure functions to assign dispersion coefficients. Finally, some approaches simply assume the suspended sediment and the velocity are uniformly distributed in the vertical and solve the vertically averaged version of eqn. 18.44 assuming that the dispersion coefficients are zero. The last approach gives approximately correct results only for low values of the Rouse number; in most cases, either of the other two is a better choice.

Table 18.1 Descriptions of eight public domain flow and morphodynamics models available in the iRIC interface.

Model	Flow Dimension		Time Variation	Hydrostatic	Coordinate System	Roughness Model	Inflow/ Outflow	Computational Time	Computational time per time-step
	2D	3D							
FaSTMECH	Y	Quasi-3D	Quasi-Steady	Y	Structured Curvilinear Orthogonal	Drag Coefficient, Roughness Height	Single/Single	7 s-1000 Iterations	0.007 s
Morpho2D	Y		Unsteady	Y	Structured General Curvilinear	Manning's *n*	Multiple/Single	17956.34 s, 3600 s simulation, 0.01dt, 360000 time-steps	0.04987 s
Nays2D	Y		Unsteady	Y	Structured General Curvilinear	Manning's *n*	Two/Single	737.1 s, 3600 s simulation, 0.1 dt, 36000 time- steps	0.0204 s
Nays2DFlood	Y		Unsteady	Y	Structured General Curvilinear	Manning's *n*	Multiple/Single	562.43 s, 3600 s simulation, 0.1dt, 36000 time-steps	
NaysCube	N	Y	Unsteady	N	Structured General Curvilinear	Manning's *n*	Single/Single	40342.05 s, 3600 s simulation, 0.05dt, 72000 time-steps	0.5603 s
Delft3D	Y	Y*	Unsteady	Y	Structured General Curvilinear		Multiple/Multiple		
SToRM	Y		Unsteady	Y	Unstructured - Triangles		Multiple/Multiple	784.04s, 3600 s Simulation, 0.1dt, 36000 time-steps	0.0218 s
River2D	Y		Unsteady	Y	Unstructured - Triangles	Roughness Height	Multiple/Multiple	4757.11 s, 7200 s simulation, variable dt 0.1 s - 100 s, 91 time-steps	52.2759 s

Model	Bed Evolution	Bank Erosion	Bed Material	Sediment-Transport Algorithm	Bedload/ Suspended Load	Turbulence Treatment	Local Grid Refinement	Vegetation Roughness	Spatially Varying Roghness	Periodic Boundary Condition
FaSTMECH	Y	N	Single Grainsize/ Mixed Grainsize	Yalin; Engelund- Hansen; Meyer- Peter Muller; Parker	Y/N	Zero Equation Model	N	N	Y	N
Morpho2D	Y	N	Single Grainsize/ Mixed Grainsize	Ashida and Michiue	Y/Y	Zero Equation Model	N	Y	Y	N
Nays2D	Y	Y	Single Grainsize/ Mixed Grainsize	Ashida and Michiue, Meyer- Peter Muller	Y/Y	Constant Equation, Zero Equaution, K-E Model	N	Y	Y	Y
Nays2DFlood	N	N	None	None Ashida and	None		N	N	Y	N
NaysCube	Y	N	Single Grainsize	Michiue, Meyer- Peter Muller, Kovacs Parker	Y/Y	Linear K-E Model, Nonlinear K-E Model	N	Y	Y	Y
Delft3D	Y*	N					N		Y	N
SToRM	N	N	None	None	None	Constant Equation, Zero Equaution	N	N	Y	N
River2D	N	N	None	None	None	Zero Equation Model	Y	N	Y	N

*Hydrostatic with morphodynamics, non-hydrostatic without morphodynamics.

Bar evolution

The primary test of any morphodynamics model is the ability of the model to reproduce simple bar forms in rivers such as alternate bars, point bars and braid bars. Some of the very first applications of two-dimensional models for flow, sediment transport and bed evolution were carried out in order to predict the formation of these simple bar types. Shimizu and Itakura (1985, 1989) showed that a two-dimensional flow model could be used to predict the formation of alternating bars in simple straight channels. They compared the equilibrium results of both computational and experimental studies of bed evolution using the experimental results of Hasegawa (1984). In the experimental case, the bed was initially flat, and in the computational case, the bed was flat with the exception of a single small perturbation. In both cases, transport was exclusively as bedload. The two-dimensional model was remarkably accurate in predicting the wavelength and amplitude of bed adjustment of alternate bars. Figure 18.2 shows the result of bed evolution in an initially flat-bedded straight channel using Nays2D in the iRIC interface.

In the same study, Shimizu and Itakura (1985, 1989) also showed that the growth and stability of point bars could be treated with a two-dimensional flow model provided that an empirical correction for the presence of secondary flows due to channel curvature was made in the computation for cross-stream sediment flux. The same methodology was employed by Struiksma (1985), who used a depth-averaged model to investigate bed evolution in the Waal River. Thus, even some of the first applications of two-dimensional flow and bed evolution models recognized that at least some three-dimensional processes had to be included in an approximate fashion to get reasonable results.

Since these early models for bed evolution were developed, many other model developers have produced similar results for basic bar forms found in rivers. This progress has led to the application of two-dimensional morphodynamics models to a variety of real problems in bed evolution in rivers.

Examples of two-dimensional model application

Figure 18.3 shows model results from a two-dimensional model used to predict the flow field for a single discharge in a reach of the Green River, Utah. As shown in the larger image on the left, the reach consists of a single long bend with a large island just downstream of the bend apex. The inset diagrams show vertically averaged flow velocity and water-surface elevation for the region of the bend near the upstream end of the island. These results were developed using the two-dimensional version of FaSTMECH in the iRIC interface; they exemplify the two-dimensional response of flow to bathymetric variation. As the flow approaches the island, the water-surface velocity increases in response to the shoaling depth, driving cross-stream flow that steers the water around the head of the island, a response typically referred to as 'topographic steering'. This response is a critical element of the improvement found in two-dimensional models relative to one-dimensional treatments. As described by Dietrich and Smith (1983), Nelson and Smith (1989b) and many others, this simple effect is a critically important part of the formation and stability of river bars. For the case depicted, researchers were interested in understanding the stability of the relatively fine deposit at the head of the island, which was known to be an important region for spawning habitat of endangered fish species.

Figure 18.4 shows the morphological evolution of a reach of the Kootenai River, Idaho, over a hydrograph during almost 3 years using the Nays2D model within the iRIC interface. For this modelling application, researchers were interested in how much fine material would collect in artificially constructed pools in this gravel-bed channel. Using the measured suspended loads for the upstream boundary conditions, the Nays2D model was used to predict the deposition of fines in the reach over multiple years. The results show that fines would fill the pools during low-flow periods, but also that these fines would be partially or completely removed during higher flows, resulting in a slow filling only of certain areas (primarily certain pools with preferential deposition), as shown in Fig. 18.4. Hence the model shows that, at least for typical hydrographs, at least some long-term deposition of fines in certain pools would occur. This information was used to refine the design morphology for this channel restoration project.

18.7 Three-dimensional models

Three-dimensional coupled models for flow, sediment transport and bed evolution have become relatively common over the last decade. Nevertheless, they are used far less frequently than their two-dimensional counterparts. This is due to the difficulty of constructing full three-dimensional flow solutions in complex domains and to computational limitations arising because bed evolution models generally require hundreds or thousands of iterative calculations for steady models or a similar number of time steps for unsteady models. Therefore, full three-dimensional coupled models are still considered to be prohibitively time consuming except for specialized small-scale calculations, such as scour near structures. However, there are so-called 'quasi-three-dimensional' or 2.5-dimensional erodible

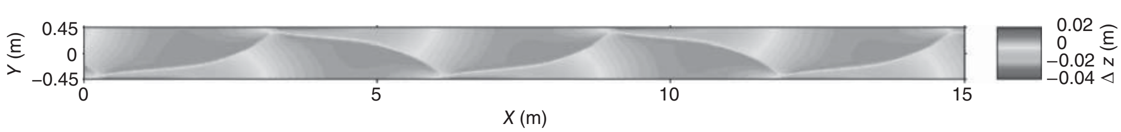


Figure 18.2 Alternate bars arising on an initially flat bed using the Nays2D morphodynamics model available in iRIC.

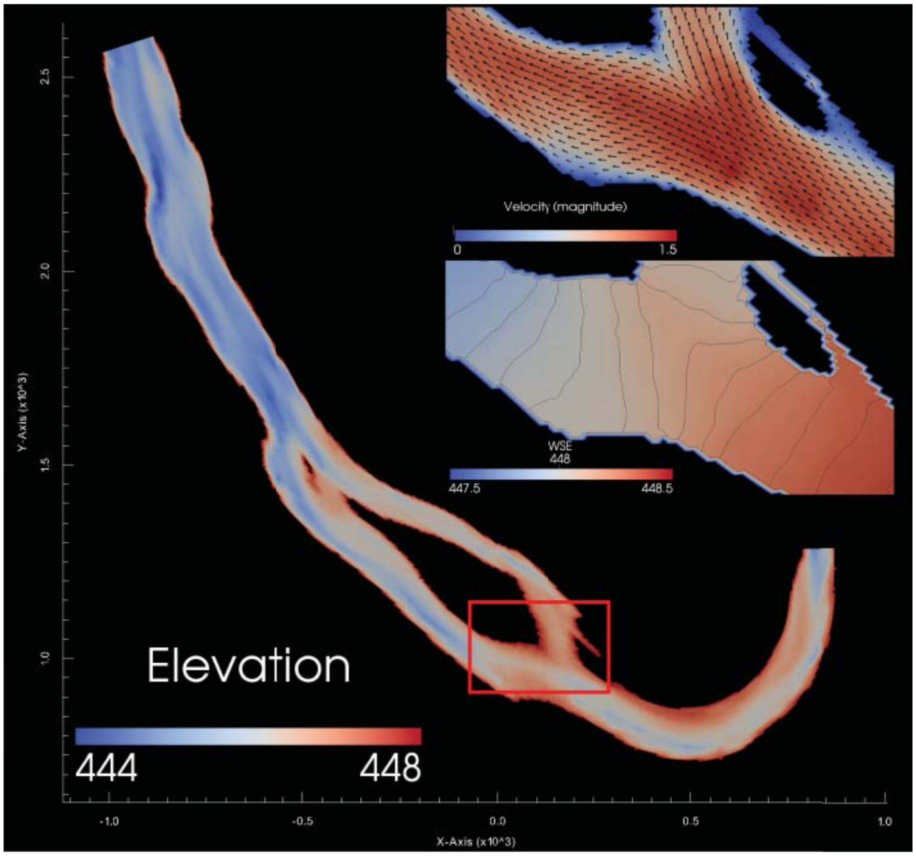


Figure 18.3 FaSTMECH model predictions for vertically averaged flow vectors (top inset) and water-surface elevation (bottom inset) on a reach of the Green River. Elevations in metres and velocity in metres per second. (*See plate section for color representation of this figure.*)

bed models that treat a good deal of the three-dimensional processes while avoiding the numerical overhead of a full three-dimensional solution, which will be discussed below. These 2.5-dimensional approaches provide a method for treating some three-dimensional processes without incurring the penalties that a fully three-dimensional model would. As computational resources continue to increase, the applicability and utility of truly three-dimensional approaches will greatly expand, with a commensurate expansion in the understanding of certain processes such as non-hydrostatic effects that are inescapably three-dimensional and cannot be captured adequately by simpler models. In this section, both quasi- and fully three-dimensional approaches will be discussed with examples.

Three-dimensional processes

In addition to the obvious improvement of predicting velocity components and stresses throughout the flow, three-dimensional models introduce three distinctly important physical processes that are not captured in one- or two-dimensional models. First, and perhaps most importantly, three-dimensional approaches allow the prediction of secondary flows. Secondary flows are defined as flows with no net discharge, acting perpendicular to the streamlines of the vertically averaged flow. The most common example is the helical flow found in meander bends, but there are others. Secondary flows are commonly driven by channel (or streamline) curvature or by gradients in normal Reynolds stress components (so-called 'turbulence-driven' secondary flows). Secondary flows produce a difference in vector direction of the flow over the flow depth. In the case of a meander bend, helical flow is produced that is directed towards the centre of curvature near the bed and away from the centre of curvature near the surface of the flow. This pattern results in a tendency for sediment deposition near the inner bank of channel bends (point bars) as discussed in fluvial geomorphology texts. Secondary flows are also responsible for many similar effects that are not so obvious and they play an important role in the evolution and stability of river bars through an interplay with topographic steering and gravitational effects on sediment transport. Except for the simple case of vanishingly low Rouse number suspension (in which secondary flows produce no net advection of sediment), secondary flows play a critically important role in determining erodible bed behaviour.

The second enhancement produced by a three-dimensional model is the precise treatment of momentum fluxes that vary in

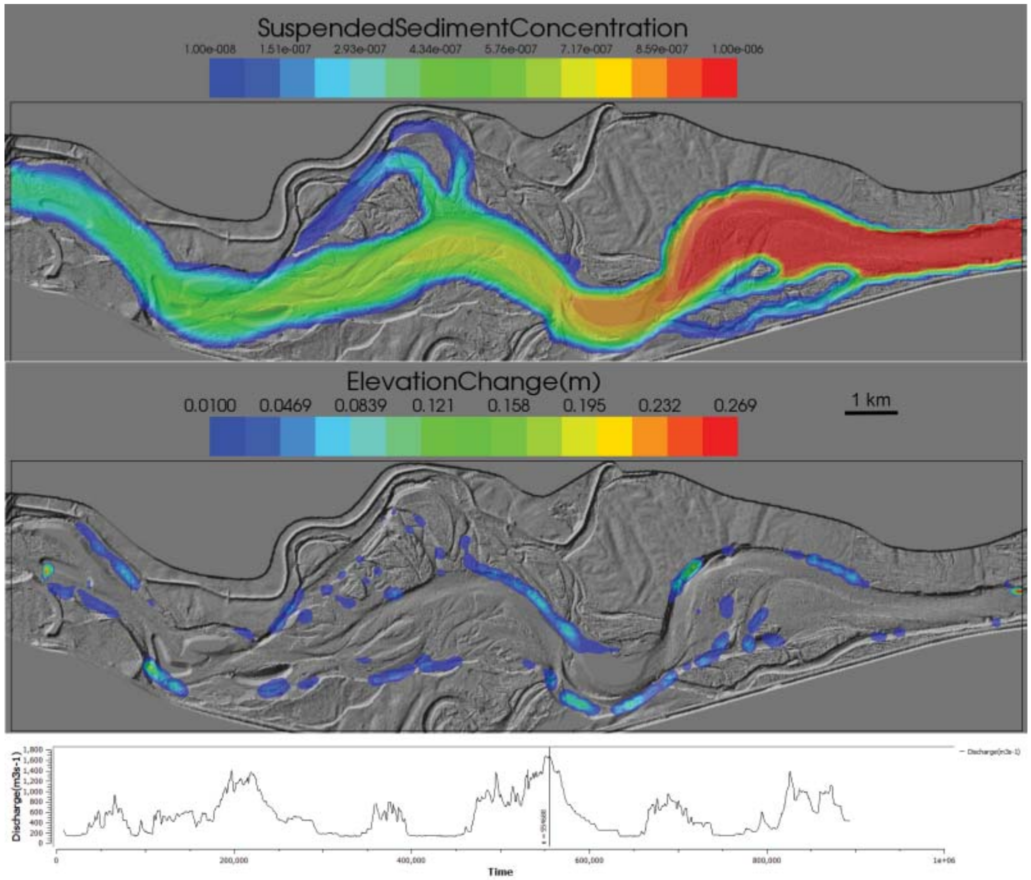


Figure 18.4 Computational prediction of the concentration and deposition thickness of fine sediment (suspended load sizes) in a reach of the Kootenai River. Time steps are 100 s, so the total time of evolution is slightly less than 3 years. The bed is shown at the time indicated by the line in the bottom diagram. (*See plate section for color representation of this figure.*)

the vertical. These are sometimes referred to as 'redistribution of momentum' effects. The helical flow discussed briefly above provides an ideal example. Because rivers tend to behave at least somewhat like simple two-dimensional boundary layer flows, it is typical for velocity to increase away from the bed. If a cross-stream helical flow is present, it will tend to advect low streamwise momentum in one direction and high streamwise momentum in another direction, resulting in a net lateral flux of streamwise momentum that cannot be predicted from the vertically averaged velocity field. This topic is directly related to the specification of dispersion coefficients, which are one way to attempt to capture the redistribution of momentum effects in a lower dimensional model. Fully three-dimensional models automatically treat these effects correctly. Redistribution of momentum effects are important for generating velocity maxima below the free surface, as shown, for example, by Shimizu *et al.* (1991), and can significantly alter the stress patterns on the channel bed and banks.

The final enhancement found in a three-dimensional model is the treatment of non-hydrostatic effects. These effects are notoriously difficult to treat in any approximate manner in a lower dimensional model and solution of the full three-dimensional equations is required. Fortunately, bed slopes in the direction of flow tend to be relatively gentle and the hydrostatic approximation is good in many situations. However, if flow is to be accurately predicted in regions of steep downstream bed slopes (e.g. over dunes, bedrock obstructions, etc.), then non-hydrostatic effects cannot be neglected. Generally, bedforms are treated parametrically in one- and two-dimensional coupled models, but if they are explicitly treated in a three-dimensional approach, accurate prediction of the flow requires a non-hydrostatic pressure distribution. The other notable situation where non-hydrostatic effects play a significant role is in flows over and around man-made structures, such as bridge piers. Coupled models for local scour near piers must incorporate non-hydrostatic effects.

Three-dimensional models

As the dimension of the model increases, the necessity to understand the approximations or assumptions that go into the model decreases and the need to spend more time on the numerical approaches increases, both in development and in computation. This is especially true if the three-dimensional model consists of a solution of eqns. 18.1 and 18.2 or a so-called direct numerical simulation. In that case, there is no need for a turbulence closure, although the grid spacing must be small enough to treat viscous dissipation of the turbulence. Needless to say, these models are generally not used to compute sediment transport and bed evolution. However, there are models that compute fully unsteady three-dimensional flow fields without the prescription of a standard turbulence closure. These models generally use a closure that is used to treat only fluctuations (and dissipation) that occur at a scale smaller than the model grid scale. These models are called large-eddy simulations, as they do compute the larger scales of the turbulence field directly, but treat the smaller ones parametrically. These models are especially promising for complex problems in flow and sediment transport.

As noted above, three-dimensional morphodynamics models can be roughly divided into two types: fully three-dimensional approaches, which solve for all three components of velocity in the momentum equations, and quasi-three-dimensional models, which solve for the horizontal components and assign vertical structure to the flow field. In the first group, progress was stimulated by a few early contributions. Delft3d was originally developed as a commercial code by Deltares and was subsequently released as public domain software in 2011. Olsen and Melaaen (1993) carried out coupled three-dimensional flow–sediment–bed evolution calculations for scour around a circular cylinder. Shimizu *et al.* (1991) used a three-dimensional model, but assumed that the flow was hydrostatic and did not compute bed deformation from the full three-dimensional model, although they did discuss some implications in that regard. Kimura *et al.* (2009) and Shimada *et al.* (2011) developed a practical three-dimensional morphodynamics model (NaysCUBE) with a stretched vertical coordinate that allows application to realistic river bathymetry. In the second group, there are several three-dimensional models that are based on a so-called 2.5-dimensional technique. In this method, the three-dimensional model solution is made up of a two-dimensional (vertically averaged) solution along with a separate computation for secondary flows and, in some cases, redistribution of momentum effects. An approach of this type was discussed in detail by Nelson and Smith (1989a,b) in the context of bed evolution calculations. Their method incorporated secondary flows generated by both channel and streamline curvature (i.e. the method predicted secondary flows even in straight channels if topographic non-uniformity was present). This method was subsequently generalized by incorporating the full vertically averaged equations of motion and a more robust numerical scheme, resulting in the FaSTMECH morphodynamics code (Nelson and McDonald 1996). Shimizu *et al.* (1991) used the same methods, but iterated on the vertically averaged solution to capture the redistribution of momentum effects and were able to show that a 2.5-dimensional approach was sufficient to capture these effects parametrically. Currently, the 2.5-dimensional approach is the most common method for three-dimensional mobile bed calculations because it captures some important three-dimensional features without requiring a full three-dimensional solution.

Three-dimensional sediment-transport models

Three-dimensional sediment-transport models are developed and applied almost identically to two-dimensional ones. Bedload is computed in the same manner, as are gravitational corrections to bedload fluxes. As full three-dimensional velocity fields are available, the advection–diffusion equation can be readily solved for the suspended sediment field, provided that the model incorporates some kind of eddy diffusivity closure or other treatment of the effects of turbulence in generating suspended sediment fluxes.

Bar evolution

As in the case of two-dimensional approaches, the first application of coupled three-dimensional models addressed the formation of simple bar forms. Nelson and Smith (1989b) used a 2.5-dimensional approach to predict the evolution and stability of point bars in curved channels and alternating bars in straight channels. Their model was cast in the channel-fitted coordinate system, but some terms in the equations were dropped due to scaling arguments. In Fig. 18.5, the evolution of both the bottom stress field and the topography is shown for a channel with a sine-generated planform shape and an initially flat bed using the FaSTMECH model within the iRIC modelling interface (the results are those of Tutorial 5 in the FaSTMECH educational materials available at www.i-ric.org). For the initial flat bed, the vector boundary shear stress has a clear component towards the inner bank, which is produced by the secondary flow. This produces a deposit on the inner bank of the bend and scour on the outer bank. As time progresses, the growth of the point bar tends to steer the flow along the inner bank outward and also produces a lateral slope which deflects sediment flux downslope. For the case shown, all transport is by bedload, so this correction has a significant impact on the pattern of sediment flux. Thus, development of the point bar is initially driven by curvature, but is stabilized by a combination of topographic steering and gravitational effects on sediment flux. The position of the point bar relative to the bend apex (i.e. the point of maximum channel curvature) is determined by the balance of cross-stream convergences of sediment, which are in-phase with the channel curvature and downstream convergences, which are out-of-phase with the channel curvature. The success of the relatively simple 2.5-dimensional coupled model for a variety of simple bar types indicated the potential of the technique. Since this early work, a variety of more general models based on the same concept have been applied in practical situations.

Examples of three-dimensional model applications

Although most of the initial efforts using three-dimensional morphodynamics models addressed the formation and mechanics of simple bar forms, practical applications followed closely. Nelson and McDonald (1996) used a 2.5-dimensional

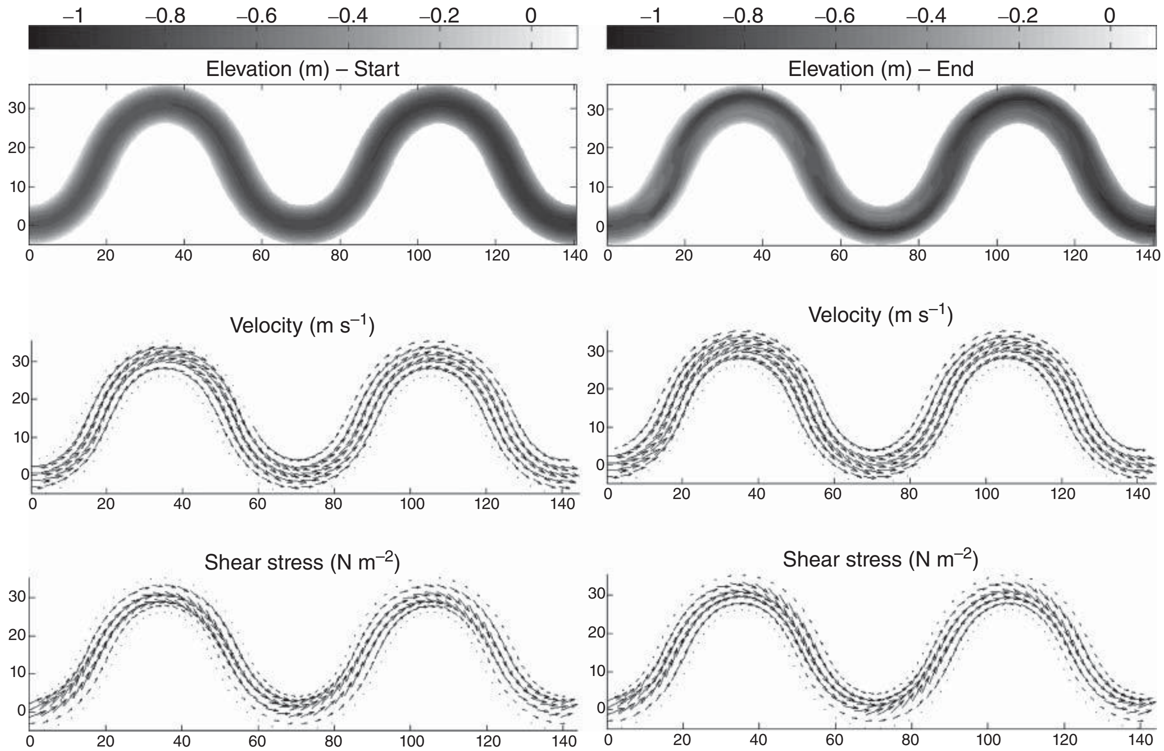


Figure 18.5 Initial evolution of a point bar using the FaSTMECH model, where the left panels correspond to the initial condition and the right panels corresponds to the evolved condition. Top panels are elevation, middle panels are velocity and bottom panels show vector bed stress. Spatial units in metres.

model to predict the details of deposition in lateral separation zones in rivers. Lateral separation zones occur in rivers and streams where bank curvature causes separation of the downstream flow from the bank, producing a region bordered by relatively slow upstream flow near the bank and by strong lateral shear along its riverward margin. These regions, also referred to as lateral separation eddies, are efficient traps of sediment and organic material and play important roles for riparian habitat and, in some cases, for recreational use (Schmidt and Graf 1990).

As already discussed in the section on two-dimensional flow modelling, a two-dimensional flow can predict the presence of lateral separation zones and can also predict deposition within them for cases when the mainstem sediment concentration is relatively high. However, because there is a separation streamline between the mainstem and the eddy region, a vertically averaged two-dimensional model can predict transport across the eddy boundary (i.e. across the reattachment streamline) only by diffusion. This is true because, by definition, there is no component of flow across the streamline joining the separation point and the reattachment point. However, for the case of bedload or suspended load with significant vertical structure, this is incorrect. Laboratory observations show that there are strong three-dimensional effects producing advection directly into the eddy. This effect is principally a product of secondary flows generated at the riverward margin of the eddy, which tends to produce flow into the eddy near the bed and out of the eddy near the surface. Thus, for lateral separation deposits formed by bedload or suspended load distributed non-uniformly in the vertical, this secondary flow creates a strong capability for capturing sediment.

Using the FaSTMECH 2.5-dimensional model, Nelson and McDonald (1996) showed using field and laboratory data that the approach could predict the secondary flows in eddies and the subsequent evolution of lateral separation bar deposits. More recently, more complete field data sets have been used to further test this approach. Figure 18.6 shows a comparison between vertically averaged velocities in a lateral separation eddy in the Colorado River in Grand Canyon compared with the predictions of the model. This approach correctly predicts the turning of the bed stress direction towards the eddy centre relative to the vertically averaged velocity direction. Hence at least some important three-dimensional effects can be predicted using a 2.5-dimensional approach at significant savings of computational effort relative to a fully three-dimensional treatment.

One of the principal uses of morphodynamics models for rivers is in evaluating the impacts of natural and anthropogenic changes in flow, sediment supply and channel form on river morphology. Figure 18.7 shows an example of the FaSTMECH-predicted evolution of bathymetry and grain size over a 60-day period of relatively high flows following the emplacement of three spur dikes (along the upper or north bank of the river downstream of the bend apex) on a meander bend in the Kootenai River in Idaho. Using similar calculations,

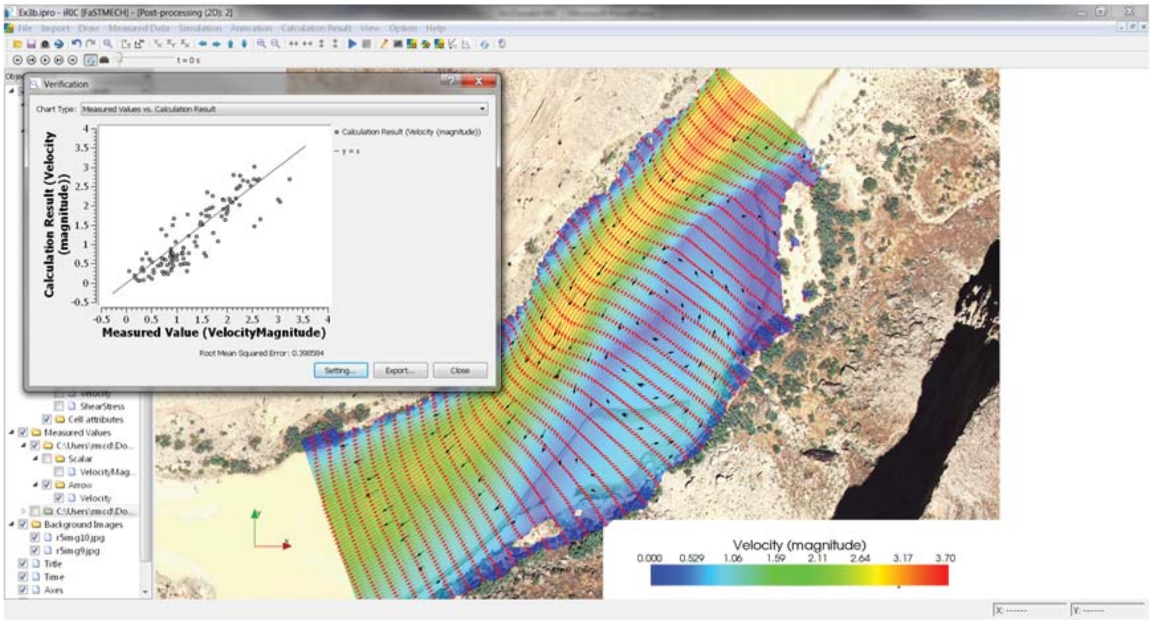


Figure 18.6 Computed vertically averaged velocity vectors compared with measured values in the Eminence Break region of the Colorado River in Grand Canyon. Calculated values are from the FaSTMECH 2.5-dimensional model in the iRIC software interface. The inset shows a comparison of the velocity magnitude between measured and calculated values. The channel is approximately 250 m wide in this view. (*See plate section for color representation of this figure.*)

morphodynamics models were used throughout a 35 km reach of the Kootenai River both pre- and post-construction to evaluate channel restoration and stabilization efforts including the modifications shown in Fig. 18.7. This included evaluating the size, spacing and number of spur dikes, and also the filling and emptying of artificial pools (as described above), the stability and infiltration of fines for emplacements of gravel spawning substrate, the stability of reconstructed back-bar channels and so forth. This kind of evaluation is essential for designing and refining channel changes proposed for meeting specific flow and morphology objectives.

Morphodynamics models can also be used to extend the evaluation of water-surface elevations and flow velocities during flood events to the evaluation of potential hazards associated with changes in bed elevation during those flooding events. Figure 18.8 shows before and after channel morphology for a reach of the Knik River in Alaska with a known issue with bridge pier undermining during flood events. Using the hydrograph in Fig. 18.8(c) and the initial topography shown in Fig. 18.8(a) along with an assumed sediment supply, Conaway (2006) used the FaSTMECH approach to predict the contraction scour in the bridge opening in the centre of the reach. The predicted post-flood bathymetry is shown in Fig. 18.8(b), with local scour in the bridge opening at or below the bridge pier footing [see Conaway (2006) for details and comparisons to bed level measurements].

The 2.5-dimensional morphodynamics examples presented above do not depend on fully three-dimensional effects to obtain the basic response of the bed, but there are problems that do require a fully three-dimensional approach to predict the appropriate bed behaviour. Figure 18.9 show results for the cross-sectional flow field and bed evolution for a simple, straight, narrow channel with an initially flat bed. These results were developed using the NaysCUBE non-hydrostatic three-dimensional flow model developed by Kimura *et al.* (2009) and Shimada *et al.* (2011) incorporating a nonlinear k–ε turbulence closure (Kimura and Hosoda 2003). For that case, the prediction of anisotropy in the turbulence field results in the generation of turbulence-driven secondary flow, as shown in Fig. 18.9(a), with the associated suppression of the velocity maximum below the surface near the walls (Fig. 18.9b) and generation of helical patterns of flow. These cells of helical flow produce convergences in the sediment transport field and longitudinal streaks in the bathymetry (Fig. 18.9c). The patterns are in excellent agreement with experimental data (see Suzuki *et al.* 2013).

A fully three-dimensional model is required for predicting bed response in situations with strongly non-hydrostatic flow (i.e. flows with strong vertical velocities and accelerations). Flow and bed evolution (local scour) near a bridge pier is a simple example where a three-dimensional model must be used for a first-principles prediction of bed change. Figure 18.10 shows bed evolution predictions for a single cylindrical pier using the NaysEddy three-dimensional flow and sediment-transport model developed by Nabi *et al.* (2013). This model has recently been added to the iRIC interface and is freely available for use. As seen in Fig. 18.10, the complex patterns of vertical flow separation and the formation of a so-called horseshoe vortex result in a realistic depiction of pier scour [in good

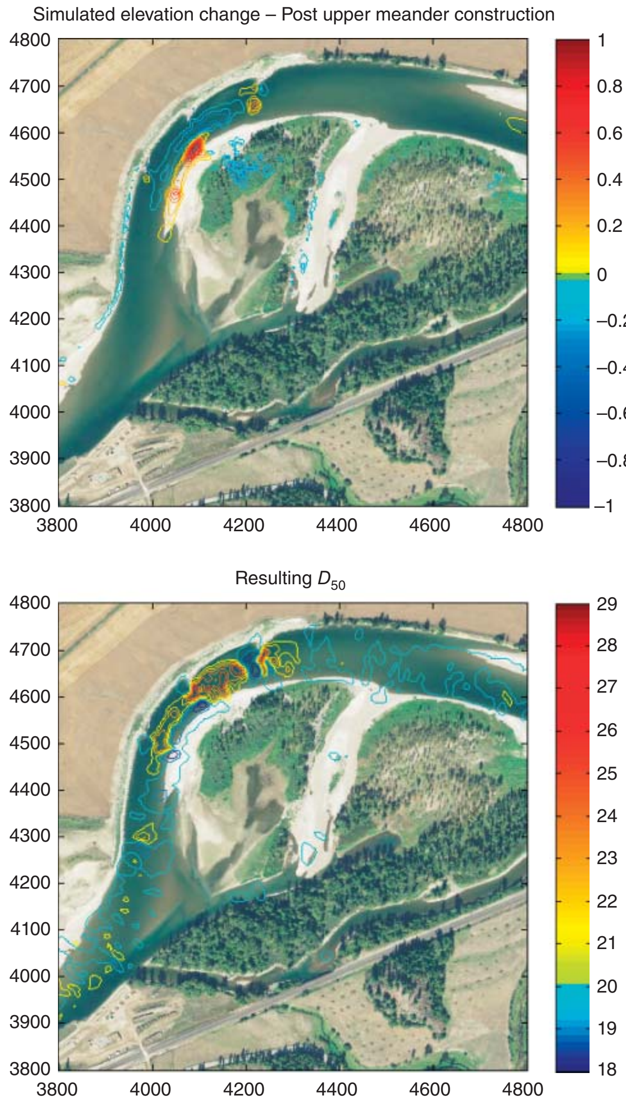


Figure 18.7 Bed morphology changes (metres) and surficial grain sizes (millimetres) predicted using FaSTMECH after the emplacement of the three spur dikes shown on the right bank. (*See plate section for color representation of this figure.*)

agreement with data; see Nabi *et al.* (2013) and Kim *et al.* (2014)]. These flow features can only be reproduced by a model with non-hydrostatic dynamics, so predicting this behaviour requires a fully three-dimensional approach.

18.8 Bank evolution models

For the most part, river morphodynamics models based on detailed flow and sediment transport modelling make predictions about the bed of the channel assuming that the banks are stable. To at least some extent, this choice has been dictated by computational restrictions, as these models typically look at change over time-scales that are short compared with those of bank evolution because it is not feasible to apply such models over longer time-scales. Despite this limitation, there have been some attempts over the past few decades to calculate the movement of banks using models that combine predicted near-bank flow properties (velocities, lateral stress, etc.) and various geotechnical properties of river banks to predict the risk of bank erosion or the potential bank erosion rate (Nagata *et al.* 2000). Some researchers have also used morphodynamics to treat situations where bed and bank erosion occur at similar time-scales, such as in some braided rivers or laboratory channel with banks composed of non-cohesive material. However, until very recently there has been little progress in using multidimensional flow and sediment-transport models to compute bank evolution over geomorphically significant time-scales for rivers where banks evolve slowly compared with channel beds, as is typically the case in rivers. Progress in understanding the fundamental time-scales of bank erosion and stabilization has recently been combined with detailed flow and sediment-transport modelling to predict the long-term evolution of a simple channel.

Asahi *et al.* (2013) used Nays2D, a two-dimensional model for flow and sediment transport including secondary flow effects, along with simple treatment for the time-scale of outer bank protection by slump blocks (Parker *et al.* 2011) and a model for inner bank stabilization by vegetation to predict the evolution of an initially straight channel with a single bend perturbation into a realistic meandering planform. The model used a time compression algorithm to reduce computational times and also incorporated a geometric treatment for treating meander cutoffs formed when channel banks meet. Figure 18.11 shows an evolution case from the initial straight channel with a single bend perturbation up to a fully formed meander belt with a single cutoff. Although calculations of this type are only just becoming available, the linkage from small-scale process up through long-term landscape evolution is impressive. More importantly, this approach bridges the gap between the statistics of channel form and relatively short time-scale parameterization of physical processes such as vegetation growth, bank cohesion and hydrology. Both the morphodynamics model and the bank evolution model for these results are freely available within the iRIC interface.

18.9 Bedform models

Computing the morphological changes in river beds and banks in response to changes in flow and sediment supply using multidimensional hydrodynamics models coupled to sediment-transport algorithms has become increasingly common for addressing practical problems in rivers. Approaches of this type typically concentrate on directly predicting relatively large-scale features, such as bars, and treat smaller-scale features, such as dunes and other bedforms, empirically through the specification of roughness and the use of simple form-drag relations. However, for certain problems, bedform fields evolve rapidly and in a complex manner during floods, requiring a

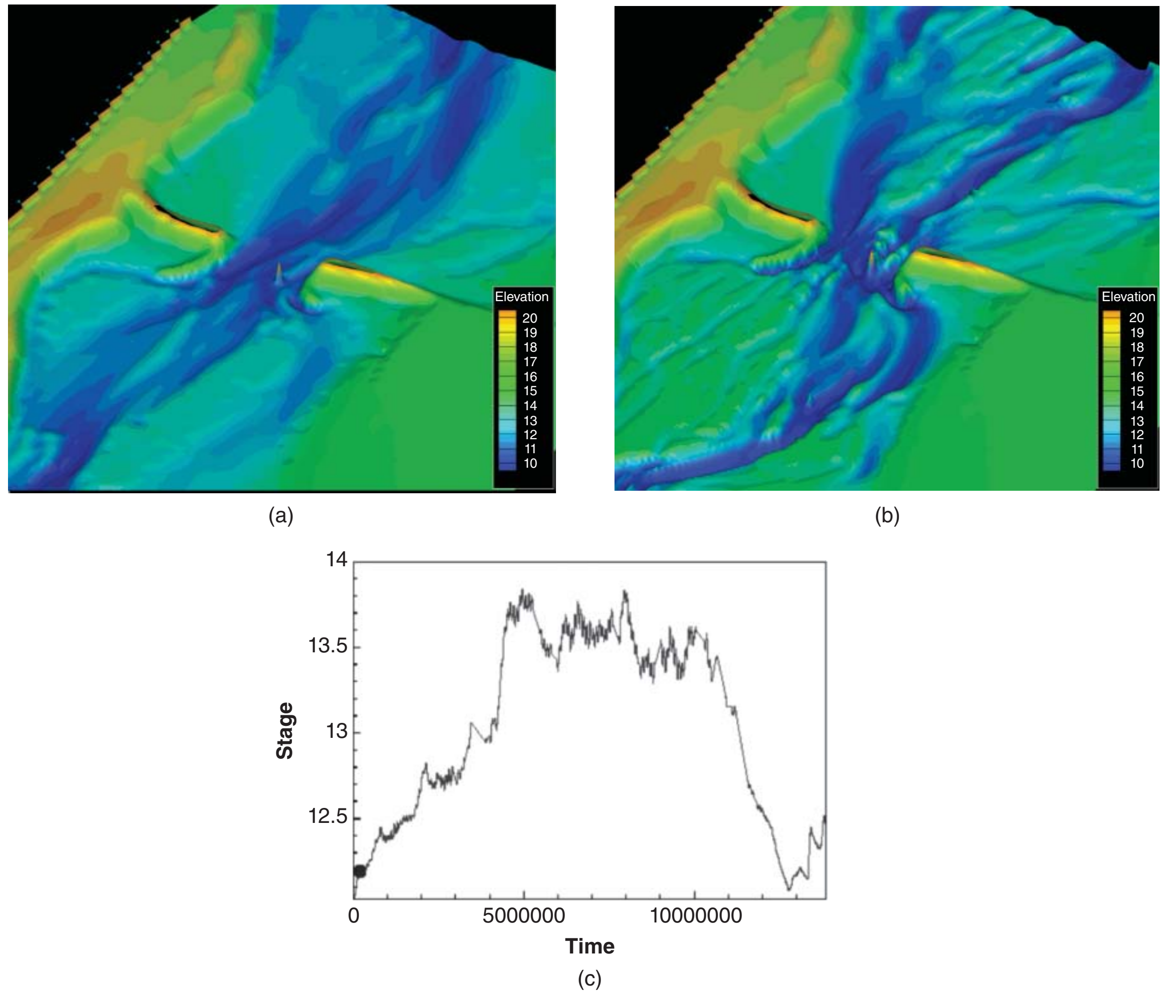


Figure 18.8 Figure showing the evolution of the Knik River channel during a high-flow event predicted using the FaSTMECH 2.5-dimensional model. (a) Pre-flood bathymetry; (b) Bathymetry after the discharge event shown in (c) (downstream flow from upper right to lower left). For scale on this perspective view, the bridge opening is ~200 m wide. Elevations and stage in metres, time in seconds. (*See plate section for color representation of this figure.*)

great deal of calibration data (i.e. field measurements of velocity and water-surface elevation) to determine the time-varying roughness and appropriate form-drag corrections for use in large-scale morphodynamics models. Unfortunately, in many cases, it is difficult or impossible to predict flow and morphological response because information about bedforms is poorly constrained or entirely unknown. Recently, small-scale morphodynamics models that predict the initiation, growth and evolution of bedforms in simple flows from first principles have been developed and tested. Based on this success, in the future it will almost certainly be plausible to develop computational models that can treat the morphodynamics of both bars and bedforms within a single, unified framework. However, at present, using models with the resolution and turbulence treatments required to treat bedform dynamics in large-scale river simulations is still computationally impractical. For the bedform models, grid resolutions on the order of a few millimetres are common, whereas scales in river morphology models are rarely less than a metre except in the very smallest streams. Nelson *et al.* (2009) described a simple technique for incorporating the results of high-resolution bedform models into larger scale models for prediction of flow and morphological evolution over long river reaches and offered an example of the method for the Kootenai River in Idaho, both to clarify the technique and to show how including dynamic bedform effects can alter model predictions.

The first bedform development models that predicted the formation, evolution and stability of bedforms on an initially flat bed using a first principles approach were described and tested by Giri and Shimizu (2006, 2007). The reader is referred to their papers for details on the modelling approach, but briefly their model is based on computational solution of the non-hydrostatic two-dimensional (vertical-streamwise) Reynolds-averaged Navier–Stokes equations using a non-linear k–ε closure. These equations are solved on a time-varying boundary-fitted coordinate system and free-surface effects are explicitly treated. Although this flow model was employed, there

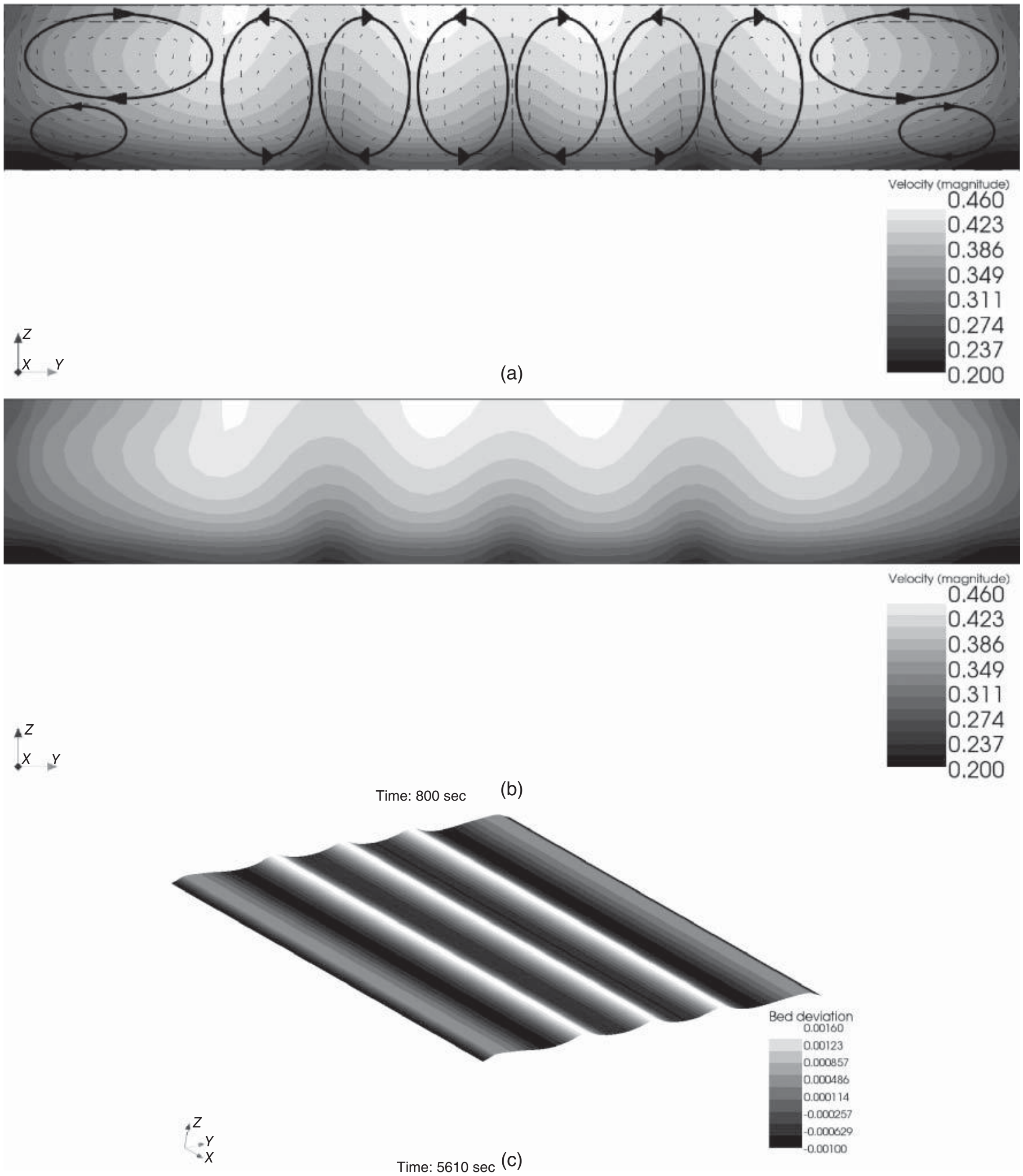


Figure 18.9 Prediction of the formation of longitudinal streaks using the three-dimensional model Nay3d. (a) Cross-stream vertical velocities on the initial flat bed; (b) streamwise velocity contours in a cross-section; (c) evolved elevations. Elevations in metres and velocities in metres per second.

are other flow prediction methods that may be equally suitable, including those described by Tjerry and Fredsoe (2005) and Niemann *et al.* (2011). To predict bedform formation and evolution, Giri and Shimizu (2006, 2007) coupled their flow model to the disequilibrium bedload model described by Nakagawa and Tsujimoto (1980) and an advection–diffusion computation for suspended load using the flux form of lower boundary conditions suggested by Itakura and Kishi (1980) and Smith and McLean (1977). The resulting model has been compared with measured velocity and pressure fields over bedforms (Nelson *et al.* 2005; Giri and Shimizu 2006) and predictions of bedform behaviour from the approach have been favourably compared with observations (Giri and Shimizu 2007; Giri *et al.* 2007; Shimizu *et al.* 2009).

The single significant shortcoming of the approach developed and tested by Giri and Shimizu (2006, 2007) was that it was limited to two-dimensional bedforms, despite the fact that many, if not most, natural bedforms have three-dimensional structure.

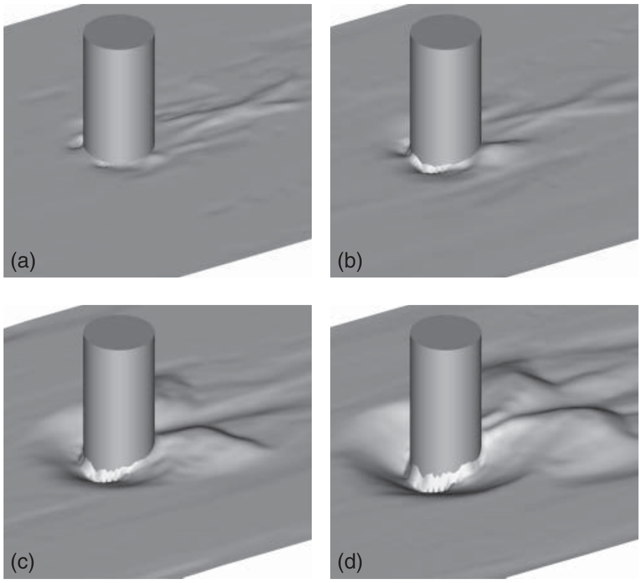


Figure 18.10 Evolution of bathymetry in unidirectional flow with a single, cylindrical pier using the NaysEddy model.

This limitation was quickly rectified by the work of Nabi *et al.* (2013), who developed a three-dimensional non-hydrostatic large-eddy flow model and a sophisticated particle-based approach for bedload and suspended load. A typical prediction from the Nabi *et al.* (2013) approach is shown in Fig. 18.12, demonstrating the evolution of a bedform field from an initially flat bed to a train of well-developed features. More than 10 years ago, when the first edition of this book and this chapter were published, the conclusions pointed out that the bedform problem was important and still unsolved. Progress may be incremental, but it is good to see that even the difficult problems are finding solutions. Although incorporating this predictive capability directly into larger scale computational models for morphodynamics in rivers is still problematic, in principle the capability is there and, as noted above, there are techniques for indirectly coupling form drag and bedform morphology from small-scale bedform models within morphodynamics models with much coarser grids (e.g. see Nelson *et al.* 2009).

18.10 Practical considerations

Choosing an appropriate model

One of the most difficult processes for a beginning modeller is choosing the appropriate model for the problem they are trying to solve. Experience makes this choice easier, but guidelines on model speed and characteristics can also be helpful. The descriptions offered above with regard to what processes are captured by models that resolve differing dimensions are also of critical value. For example, if the modeller knows that non-hydrostatic effects are an important part of the problem, the choices narrow dramatically. If the model needs to make predictions over very long periods, practical issues with regard to computational run times and data storage become paramount. Although a complete catalogue of model characteristics and speeds are outside the scope of this chapter, Table 18.1 does give that information for the free, public domain models available in iRIC that treat two- and three-dimensional problems. All of these models are accompanied by detailed information and tutorials available on the iRIC website at www.i-ric.org.

For each model, Table 18.1 gives information about the dimensions of the model, the coordinate systems (both structured and unstructured), roughness parameterization, the available sediment-transport treatments and so forth. All this information is also available on the iRIC website in far more detail. In addition to information on model characteristics, each of the models was used to compute flow (and, where available, sediment transport and bed deformation) for a reach of the Green River over a short period using a single value of discharge (a steady flow). The reach used is shown in Fig. 18.3 and consists of a simple, mostly single-thread, channel bend with an island about half way through the reach of interest. The actual application for this reach was a study to understand the effects of sediment transport on fish spawning habitat, but here this reach serves only to compare the models. For each model, a grid or mesh with approximately 20,000 nodes was employed. For the unsteady models (all but FaSTMECH), the time step was chosen as large as possible while still maintaining model stability. The planform discretization was set to about 7 m (corresponding to an area of 49 m^2 for a square element) for the structured

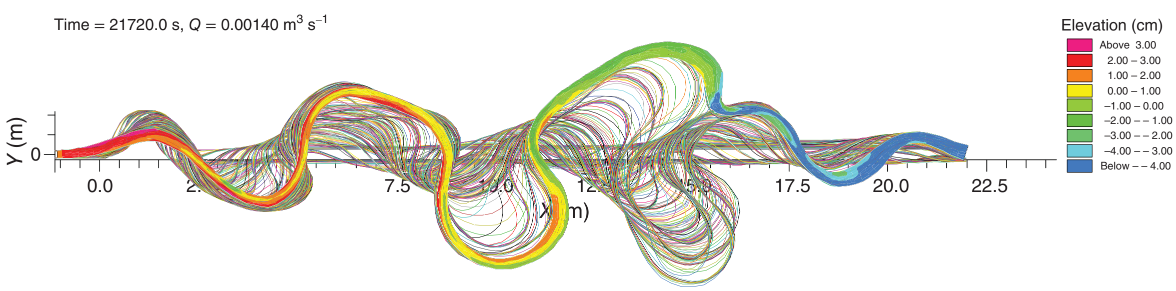


Figure 18.11 Channel planform evolution predicted starting with a straight channel with a single bend perturbation. Differently shaded lines correspond to the position of the channel at different times. Note the cutoff present near the downstream end of the reach. (*See plate section for color representation of this figure.*)

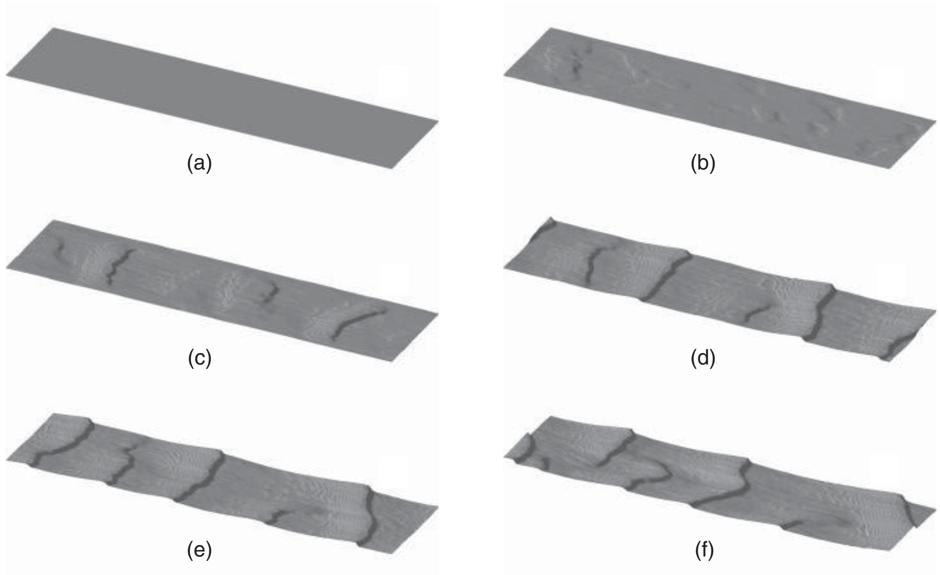


Figure 18.12 Evolution of bedforms on an initially flat bed using NaysEddy. Alphabetical progression from (a) to (f) on the panels corresponds to time progression in the calculation.

grids using the same orthogonal curvilinear grid and to an average element area of 50 m^2 for the unstructured grid models. Note that this yields more total grid points for the structured grid models owing to their inability to fit channel boundaries closely. Typically, the structured grid models used a total of about 25,000 and the unstructured grids used about 15,000 grid points. For the two models with three-dimensional capabilities (FaSTMECH and NaysCUBE), the horizontal grid resolution was set from the grid for the two-dimensional models and the vertical dimension was split into 11 unevenly spaced levels.

As shown in the column of Table 18.1 headed ‘Computational time’, the requirements for computer time (running on a PC laptop with a single 3.2 GHz Intel I7 processor; any use of trade, firm or product names is for descriptive purposes only and does not imply endorsement by the US Government) vary from 7 s for FaSTMECH to over 40,000 s for NaysCUBE, a difference of almost four orders of magnitude. This highlights the balance between low computational time and physical accuracy. The fully three-dimensional model NaysCUBE captures physical effects that the quasi-three-dimensional model FaSTMECH does not, including advection of turbulence, turbulence-driven secondary flows, non-hydrostatic effects in steep regions of the bed and so forth, but it does so at great computational expense. In some sense, the ‘art’ of modelling is striking the best possible balance between resolution of the processes and minimization of computational effort; understanding these tradeoffs is critically important. Similarly, although the unstructured grids offer great flexibility in fitting domains and allowing complex boundary conditions, results from unsteady two-dimensional structured grid models consume almost an order of magnitude less computational time. Again, if the domain of interest requires the advantages of an unstructured grid model (such as easy treatment of complex inflow and outflows) that approach is the better choice, but in many situations not requiring these specific features, a structured model that requires far less computational effort will provide similar results. Users of computational models should consider this kind of information carefully in choosing modelling schemes – there is no one single model that is always the best choice, which is really the concept behind making all the models listed in Table 18.1 available within the iRIC interface.

The modelling process

Once the user has chosen an appropriate model and assuming that the appropriate field data are available, the process of modelling flow and morphodynamics can begin. Ideally, long before this point the user will have carefully considered the resolution of topographic, bathymetric, water-surface, sediment and various other kinds of data that will be needed to solve the modelling problem. The most common reason for failure or poor performance of modelling projects is inappropriate or incomplete field data. For example, if bathymetric data are only measured on a 10 m grid and the question at hand is fish habitat prediction on a 1 m grid, the approach is flawed. The data driving any flow or sediment model must be spatially and temporally resolved at the scales of the questions to be answered. This may seem obvious, but even experienced modellers frequently fail to think carefully about the importance of field data and how the data relate to model predictions. A good rule-of-thumb is that a well-designed, carefully planned river modelling project typically consists of 90% effort in the field collecting and verifying appropriate data and about 10% effort on actually developing the model results. The modelling process itself proceeds along a path of (1) preparing field data, (2) developing a coordinate system (structured or unstructured grid depending on the model) and mapping the measured data on to the nodes of that system, (3) setting initial/boundary conditions and running initial scenarios, (4) calibration as necessary, (5) running full modelling scenarios, (6) creating model results in tabular and graphical form and (7) verifying the model to the extent allowed by available data. Each of these steps is briefly described in the following:

1 The basic data required for a flow model of a river reach are topography and bathymetry measurements that characterize the range of flows to be modelled, plus boundary and initial conditions for the model, roughness and the discharge or hydrograph to be modelled. If flow and morphodynamic modelling is to be carried out, to these data one must add detailed information on sediment grain sizes (often by size class and typically both surface and subsurface), sediment supply for the range of modelled discharges, settling velocities for the sediments of interest, bedform geometry and coverage and potentially information about sediment cohesivity. Frequently, elevation data sets come from different sources, so part of the preparation of a complete data set of elevation may include combining LiDAR, single- or multi-beam acoustic and wading measurements of elevation. Careful attention regarding a consistent choice of datum is critical, as is the use of interpolation or other refinement methods to make sure the bathymetric data set is the best possible representation of the system. All data should be georeferenced to the same coordinate system.

2 Once the basic data are in a consistent format, the user can create (graphically on the computer screen using a mouse, typically in the iRIC interface) an appropriate coordinate system for the model of choice. Practical grids only rarely exceed one million nodes in the planform domain, although some of the iRIC models have been used on significantly larger meshes or grids. After creating and, in some cases, refining that system, the user should use any of a wide variety of techniques to map the measured data on to the coordinate system. This process is not always straightforward and the mapping method should be chosen to create the best possible realization of the bathymetry at the resolution of the coordinate system. This may involve additional interpolation using aerial or satellite photographs, development of break lines for triangular interpolation networks or point-by-point hand work to develop the best possible mapping. The iRIC interface has a number of tools for this process and also tutorials demonstrating their use.

3 Once the field data have been mapped on to the modelling coordinate system, the next step is choosing reference flows (typically the flows or hydrographs that were measured during the field programme) and setting initial conditions and boundary conditions for those flows. Ideally, these flows will bracket the flows of interest for the model predictions, but in some cases this is difficult, especially if one is interested in modelling rare, extremely large flows. The larger the range of flows measured, the greater is the faith in the model and its ability to interpolate and extrapolate to other hydrographs and sediment supply conditions. After choosing the reference flows and their initial/boundary conditions, simple model runs are completed for these flows.

4 With model results for the reference discharges at which data were measured, the user can compare water-surface elevation predictions with measurements and iteratively correct roughness values and spatial distributions as required to improve the comparison. Figure 18.13 shows a comparison between measured and predicted water-surface elevations using three different constant values of drag coefficient and one spatially variable value of drag coefficient for the Green River case shown in Fig. 18.3. There is a clear minimum on the root mean square error between predicted and measured water-surface elevations as a function of drag coefficient, even using a spatially constant value. For more complex spatially varying roughness distributions involving sediment grain sizes, the error can be reduced even further. Even for the case of variations in grain size, vegetation or other spatially varying effects, this basic method can be used to calibrate roughness. Similarly, measured velocities can be used to calibrate lateral diffusivity in a two-dimensional model, measured sediment transport rates can be used to calibrate or choose sediment relations, and so forth. Although the specific parameters that require calibration will vary for different models, this calibration step should always occur if data are available, as it ensures that the model can faithfully reproduce the reference flow data, leading to greater confidence in applications of the model to flows and hydrographs other than the reference values. In situations where the necessary data are not available, literature or other approximate methods for setting roughness or other parameters can be used, but the model results must be interpreted in light of that uncertainty. Many parameters, including roughness, require calibration over a range of discharges or a suite of hydrographs, so again the importance of a wide range of reference flows is clear. In cases where bedform fields change dramatically, a bedform model such as the two discussed above should also be employed. All the modelling applications within the iRIC interface allow this calibration process to be carried out in a simple manner with graphical guidance.

5 After calibration, the model runs to treat the problem at hand can be completed. The spatial and temporal resolution of these runs will be dictated by the nature of the questions to be answered, as will the choice of the model, as noted above. For situations where many computational realizations need to be run, multiple core computers can be used to great advantage. Run times can often be shortened by using earlier runs as initial conditions for later runs (i.e. using a so-called ‘hot start’).

6 With model results in hand, preparing graphics and tabular data from the model is straightforward. In the case of the iRIC interface, there are many standard tools for one-, two- and three-dimensional visualization, including static and animated graphics for essentially all the input and output variables of the model. Many of the figures showing model results in this paper were generated within iRIC. Three-dimensional results can be cut with cross-sections to visualize vertical and cross-stream velocities and model results can also be displayed in time-series format. In addition to these standard tools, all model outputs are available in standard binary formats (CGNS) that allow direct importation of complete data sets into other visualization tools such as Matlab, Tecplot and so forth.

7 The final step before a model project is done and the results are reported is verification of the model by comparison with measured data *not* used in the calibration process. This may consist of detailed water-surface elevations, acoustic–Doppler velocity data, measurement of observed channel change or something more specific to the problem addressed by the project, such as fish habitat or another subsidiary consideration that model results are used to address. In the iRIC interface, virtually any georeferenced information can be imported into the modelling project and displayed with collocated model predictions, making this process simple, but the user can also import data into spreadsheets, other analysis programs, GIS platforms or whatever other software is available that will facilitate the appropriate comparisons.

The most important step is to begin. When the first edition of this book was published, over 10 years ago, the availability and ease-of-use of models for geomorphic change in rivers were limited. As these tools have been refined, they have also entered a stage where they can be readily applied by non-specialists, provided that care is taken in following the steps above. Users can ask (and answer!) questions that would have been outside modelling capabilities even a decade ago. With the advent of freeware and the trend towards making models open-source and in the public domain, allowing easy modification and addition of new submodels, fluvial geomorphology has really entered a new, more quantitative phase, promising both greater understanding of basic process and practical predictive capabilities for river response to environmental changes.

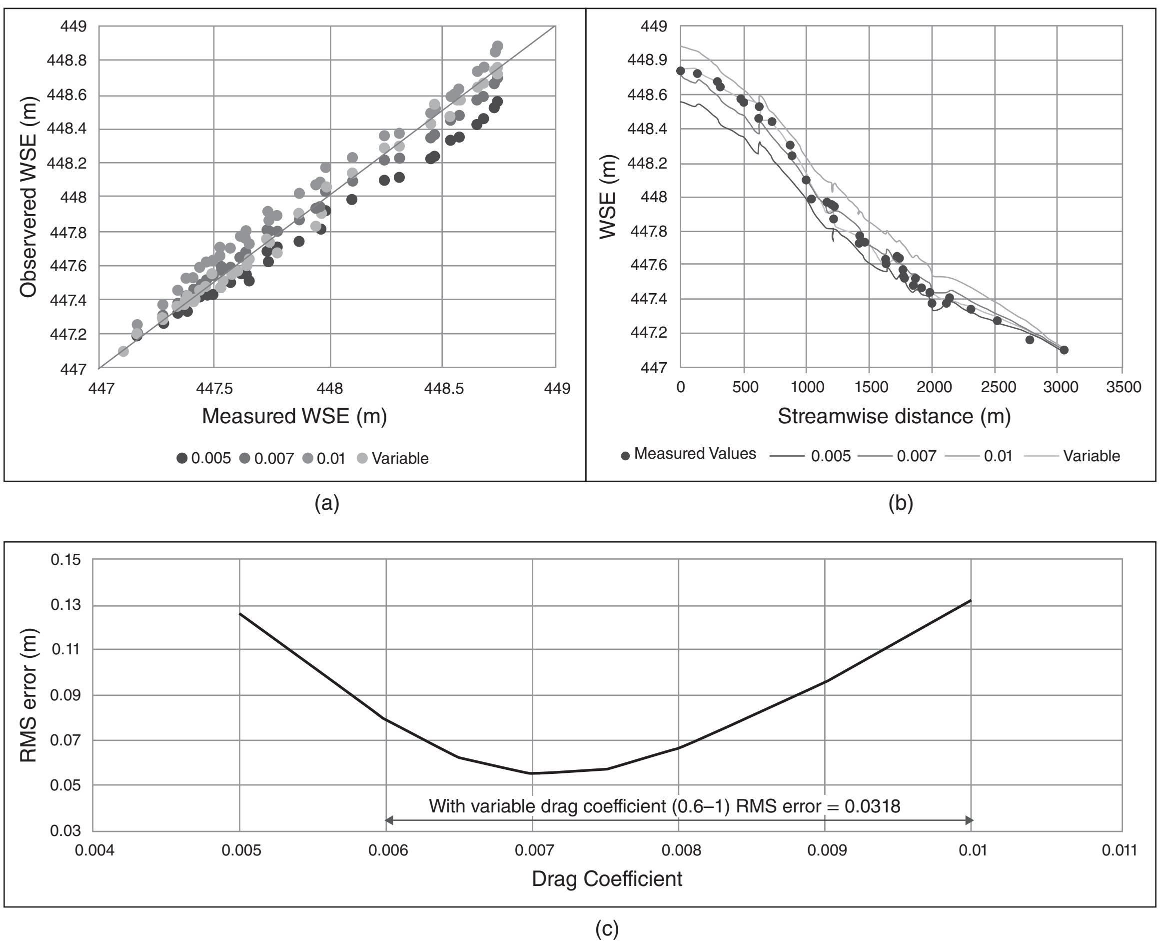


Figure 18.13 Illustration of the calibration process. (a) Comparisons between measured and predicted water-surface elevation through the reach of the Green River shown in Fig. 18.3 for three constant values of drag coefficient and one spatially value mapped by grain size. (b) Measured and predicted longitudinal profiles of water-surface elevation for the same choices of roughness. (c) Variation of error as a function of constant drag coefficient and the (lower) error associated with spatially varying roughness mapped by grain size.

18.11 Conclusions and future directions

Computational modelling of flow, sediment transport and bed evolution has made dramatic progress over the past 20 years. Over this period, most advances have been dictated by improvements in the ability to predict complex flows. In other words, the flow computation has regulated progress. However, at this point, as truly complete flow models are beginning to become available, it appears likely that research will return to the details of the sediment-transport process and how one should model it, especially in complex flows. Currently, most techniques for predicting sediment motion are extensions of methods that are strictly valid only for steady uniform flows. For example, parameterizing the forces on sediment particles that lead to bedload motion in terms of boundary shear stress assumes that all the local variability near the bed can be captured in the boundary shear stress. Generally, this is not true in complex flows and bedload motion can occur even where the mean boundary shear stress is zero. The particle-based approaches described by Nabi *et al.* (2013) and Schmeeckle (2014) appear to be particularly promising and reproduce certain physical effects that simpler methods cannot.

Perhaps more importantly for the purposes at hand, tools for predicting the behaviour of fluvial flows and their channels are now readily available for use by non-specialists. This presents a tremendous opportunity for extending qualitative understanding of processes developed over the past century into quantitative, specific predictions of river behaviour. Geomorphologists now have the capability to make meaningful linkages between short time-scale processes and long-term morphology, which offers the ability to go beyond describing what something looks like to answer questions about why it looks that way and what historical processes have been integrated over time to produce the observed channel or its deposit.

References

Abbott, M.B. and Minns, A.W. 1998. *Computational Hydraulics*, 2nd edn., Farnham: Ashgate Publishing, 557pp.

Asahi, K., Shimizu, Y., Nelson, J. and Parker, G. 2013. Numerical simulation of river meandering with self-evolving banks. *Journal of Geophysical Research, Earth Surface* 118: 2208–2229.

Chaudhry, M.H. 1993. *Open-Channel Flow*, Englewood Cliffs, NJ: Prentice Hall, 483pp.

Conaway, J.S. 2006. Comparison of long-term streambed scour data with modeled values at the Knik River, Alaska. In: *Proceedings of the Third International Conference on Scour and Erosion, Amsterdam*, Gouda: CURNET, pp. 146–153.

DeLong, L.L., Thompson, D.B. and Lee, J.K. 1997. *The Computer Program FourPt (Version 95.01) – a Model for Simulating One-dimensional, Unsteady, Open-channel Flow*, US Geological Survey Water-Resources Investigations Report 97-4016, Reston, VA: US Geological Survey, 69pp.

Dietrich, W.E. and Smith, J.D. 1983. Influence of the point bar on flow through curved channels. *Water Resources Research* 19(5): 1173–1192.

Engelund, F. 1974. Flow and bed topography in channel bends. *Journal of the Hydraulics Division, American Society of Civil Engineers* 100(HY11): 1631–1648.

Engelund, F. and Hansen, E. 1967. *A Monograph on Sediment Transport in Alluvial Streams*, Copenhagen: Technical University of Denmark, 63pp.

Fan, S. 1993. In: *Proceedings, the Second Bilateral Workshop on Understanding Sedimentation Processes and Model Evaluation*, Washington, DC: US Federal Energy Regulatory Commission, 2 vols, 747pp.

Franz, D.D. and Melching, C.S. 1997. *Full Equations (FEQ) Model for the Solution of the Full, Dynamic Equations of Motion for One-dimensional Unsteady Flow in Open Channels and Through Control Structures*, US Geological Survey Water-Resources Investigations Report 96-4240, Urbana, IL: US Geological Survey, 258pp.

Garcia, M. 2008. Sediment transport and morphodynamics. In: Garcia, M., ed., *Review of Sedimentation Engineering – Processes, Measurements, Modeling, and Practice*, ASCE Manuals and Reports on Engineering Practice No. 110, Reston, VA: American Society of Civil Engineers, pp. 21–163.

Garcia, M. and Parker, G. 1991. Entrainment of bed sediment into suspension. *Journal of Hydraulic Engineering* 117(4): 414–435.

Giri, S., Yamaguchi, S., Shimizu, Y. and Nelson, J.M. 2007. Simulating temporal response of bedform characteristics to varying flows. In: Dohmen-Janssen, C. and Hulscher, S., eds., *River, Coastal and Estuarine Morphodynamics, Proceedings of the 5th IAHR Symposium on River, Coastal and Estuarine Morphodynamics*, Boca Raton, FL: CRC Press, pp. 939–947.

Giri, S. and Shimizu, Y. 2006. Numerical computation of sand dune migration with free surface. *Water Resources Research* 42: W10422.

Giri, S. and Shimizu, Y. 2007. Validation of a numerical model for flow and bedform dynamics. *Annual Journal of Hydraulic Engineering, JSCE* 51: 139144.

Gomez, B. 1991. Bedload transport. *Earth-Science Reviews* 31: 89–132.

Hasegawa, K. 1984. *Hydraulic Research on Planimetric Forms, Bed Topographies and Flow in Alluvial Channels*, PhD Dissertation, Hokkaido University, Sapporo.

Itakura, T. and Kishi, T. 1980. Open channel flow with suspended sediments. *Journal of the Hydraulics Division, American Society of Civil Engineers* 106: 1325–1343.

Kikkawa, H., Ikeda, S. and Kitagawa, A. 1976. Flow and bed topography in curved open channels. *Journal of the Hydraulics Division, American Society of Civil Engineers* 102(HY9): 1327–1342.

Kim, H.S., Nabi, M., Kimura, I. and Shimizu, Y. 2014. Numerical investigation of local scour at two adjacent cylinders. *Advances in Water Resources* 70: 131–147.

Kimura, I. and Hosoda, T. 2003. A nonlinear k–ε model with realizability for prediction of flows around bluff bodies. *International Journal for Numerical Methods in Fluids* 42: 813–837.

Kimura, I., Uijttwaal, W.S.J., Hosoda, T. and Ali, W.S. 2009. URANS computations of shallow grid turbulence. *Journal of Hydraulic Engineering* 135(2): 118–131.

Long, C.E., Wiberg, P.L. and Nowell, A.R.M. 1993. Evaluation of von Karman's constant from integral flow parameters. *Journal of Hydraulic Engineering* 119(10): 1182–1190.

McLean, S.R. 1992. On the calculation of suspended load for noncohesive sediments, *Journal of Geophysical Research* 97(C4): 5759–5770.

Meyer-Peter, E. and Müller, R. 1948. Formulas for bed-load transport. In: *Proceedings of the 2nd Meeting, IAHR, Stockholm*, pp. 39–64.

Middleton, G.V. and Southard, J.B. 1984. *Mechanics of Sediment Movement, SEPM Short Course*, Tulsa, OK: Society of Economic Paleontologists and Mineralogists, 401pp.

Nabi, M., de Vriend, H.J., Mosselman, E., Sloff, C.J. and Shimizu, Y. 2013. Detailed simulation of morphodynamics: 3. Ripples and dunes. *Water Resources Research* 49: 5930–5943.

Nagata, N., Hosoda, T. and Muramoto, Y. 2000. Numerical analysis of river channel processes with bank erosion. *Journal of Hydraulic Engineering* 126(4): 243–252.

Nakagawa, H. and Tsujimoto, T. 1980. Sand bed instability due to bedload motion. *Journal of the Hydraulics Division, American Society of Civil Engineers* 106: 2029–2051.

Nelson, J.M. 1990. The initial instability and finite-amplitude stability of alternate bars in straight channels. *Earth-Science Reviews* 29: 97–115.

Nelson, J.M. 1996. Predictive techniques for river channel evolution and maintenance. *Water, Air and Soil Pollution* 90: 321–333.

Nelson, J.M. and Smith, J.D. 1989a. Flow in meandering channels with natural topography. In: Ikeda, S. and Parker, G., eds., River Meandering, *AGU Water Resources Monograph 12*, Washington, DC: American Geophysical Union, pp. 69–102.

Nelson, J.M. and Smith, J.D. 1989b. Evolution and stability of erodible channel beds. In: Ikeda, S. and Parker, G., eds., *River Meandering*, AGU Water Resources Monograph 12, Washington, DC: American Geophysical Union, pp. 321–377.

Nelson, J.M. and McDonald, R.R. 1996. *Mechanics and Modeling of Flow and Bed Evolution in Lateral Separation Eddies*, Glen Canyon Environmental Studies Report, 69pp.

Nelson, J.M., McLean, S.R. and Wolfe, S.R. 1993. Mean flow and turbulence fields over two-dimensional bedforms. *Water Resources Research* 29(12): 3935–3954.

Nelson, J.M., Burman, A.R., Shimizu,Y., McLean, S.R., Shreve, R.L. and Schmeeckle, M.W. 2005. Computing flow and sediment transport over bedforms. In: Parker, G. and Garcia, M., eds., *River, Coastal and Estuarine Morphodynamics, Proceedings of the 4th IAHR Symposium on River, Coastal and Estuarine Morphodynamics*, Boca Raton, FL: CRC Press, pp. 861–872.

Nelson, J.M., Shimizu, Y., McDonald, R.M. and Logan, B. 2009. Integrating computational models for bedform development with reach-scale flow and morphodynamics models for rivers. In: Vionnet, C.A., Garcia, M.H., Latrubesse, E.M. and Perillo, G.M.E., eds., *River, Coastal and Estuarine Morphodynamics, Proceedings of the 6th IAHR Symposium on River, Coastal and Estuarine Morphodynamics*, Leiden: CRC Press/Balkema, pp. 709–714.

Niemann, S.L., Fredsoe, J. and Jacobsen, N.G. 2011. Sand dunes in steady flow at low Froude numbers: dune height evolution and flow resistance. *Journal of Hydraulic Engineering* 137(1): 5–14.

Oliveira, A., Fortunato, A.B. and Baptista, A.M. 2000. Mass balance in Eulerian–Lagrangian transport simulations in estuaries. *Journal of Hydraulic Engineering* 126(8): 605–614.

Olsen, N.R.B. and Melaaen, M.C. 1993. Three-dimensional calculation of scour around cylinders. *Journal of Hydraulic Engineering* 119(9): 1048–1054.

Parker, G. 1984. Discussion of: Lateral bedload transport on side slopes (by S. Ikeda, Nov. 1982). *Journal of Hydraulic Engineering* 110(2): 197–203.

Parker, G. 2008. Transport of gravel and sediment mixtures. In: Garcia, M, ed., *Review of Sedimentation Engineering – Processes, Measurements, Modeling, and Practice*, ASCE Manuals and Reports on Engineering Practice No. 110, Reston, VA: American Society of Civil Engineers, pp. 165–251.

Parker, G., Shimizu, Y., Wilkerson, G.V., Eke, E.C., Abad, J.D., Lauer, J.W., Paola, C., Dietrich, W.E. and Voller, V.R. 2011. A new framework for modeling the migration of meandering rivers. *Earth Surface Processes Landforms* 36(1): 70–86.

Patankar, S.V. 1980. *Numerical Heat Transfer and Fluid Flow*, Washington, DC: Hemisphere, 197pp.

Press, W.H., Flannery, B.P., Teukolsky, S.A. and Vetterling, W.T. 1986. *Numerical Recipes: the Art of Scientific Computing*, 1st edn., New York: Cambridge University Press.

Rattray, M. Jr, and Mitsuda, E. 1974. Theoretical analysis of conditions in a salt wedge. *Estuarine and Coastal Marine Science* 2: 375–394.

Recking, A. 2010. A comparison between flume and field bed load transport data and consequences for surface-based bed load transport prediction. *Water Resources Research* 46(3): W03518.

Rodi, W. 1993. *Turbulence Models and Their Application in Hydraulics – A State of the Art Review*, 3rd edn., Delft: International Association for Hydraulic Research, 104pp.

Schaffranek, R.W. 1987. *Flow Model for Open-channel Reach or Network*, US Geological Survey Professional Paper 1384. Washington, DC: US Government Printing Office, 12pp.

Schmeeckle, M.W. 2014. Numerical simulation of turbulence and sediment transport of medium sand. *Journal of Geophysical Research, Earth Surface* 119: 1240–1262.

Schmidt, J. and Graf, J.B. 1990. *Aggradation and Degradation of Alluvial Sand Deposits, 1965 to 1986, Colorado River, Grand Canyon National Park, Arizona*, US Geological Survey Professional Paper 1493, Washington, DC: US Government Printing Office.

Shimada, R., Kimura, I. and Shimizu, Y. 2011. Study on 3D flow structures and bed deformations in curved open channels. *River, Coastal and Estuarine Morphodynamics*, Beijing: Tsinghua University Press, pp. 1658–1670.

Shimizu, Y. and Itakura, T. 1985. *Practical Computation of two-dimensional Flow and Bed Deformation in Alluvial Streams*, Civil Engineering Research Institute Report, Sapporo: Hokkaido Development Bureau.

Shimizu, Y. and Itakura, T. 1989. Calculation of bed variation in alluvial channels. *Journal of Hydraulic Engineering* 115(3): 367–384.

Shimizu, Y., Yamaguchi, H. and Itakura, T. 1991. Three-dimensional computation of flow and bed deformation. *Journal of Hydraulic Engineering* 116(9): 1090–1108.

Shimizu, Y., Giri, S., Yamaguchi, S. and Nelson, J.M. 2009. Numerical simulation of dune-flat bed transition and stage-discharge relationship with hysteresis. *Water Resources Research*, 45: W04429.

Smith, J.D. and McLean, S.R. 1977. Spatially averaged flow over a wavy surface. *Journal of Geophysical Research* 84(12): 1735–1746.

Smith, J.D. and McLean, S.R. 1984. A model for flow in meandering streams. *Water Resources Research* 20(9): 1301–1315.

Struiksma, N. 1985. Prediction of 2-D bed topography in rivers, *Journal of Hydraulic Engineering* 111(8): 1169–1182.

Suzuki, R., Kimura, I., Shimizu, Y. and Iwasaki, T. 2013. Numerical study on secondary currents of the second kind in wide shallow open channel flows. In: *Proceedings of Riverflow* 2013, pp. 1–8.

Tennekes, H. and Lumley, J.L. 1972. *A First Course in Turbulence*, Cambridge, MA: MIT Press, 300pp.

Tritton, D.J. 1977. *Physical Fluid Dynamics*, New York: Van Nostrand Reinhold, 362pp.

Thomas, W.A. and Chang, H. 2008. Computational modeling of sedimentation processes. In: Garcia, M., ed., *Review of Sedimentation Engineering – Processes, Measurements, Modeling, and Practice*, ASCE Manuals and Reports on Engineering Practice No. 110, Reston, VA: American Society of Civil Engineers, pp. 649–681.

Tjerry, S. and Fredsøe, J. 2005. Calculation of dune morphology. *Journal of Geophysical Research* 110: F04013.

US Army Corps of Engineers. 1993. *HEC-6, Scour and Deposition in Rivers and Reservoirs, User's Manual*, Davis, CA: Hydrologic Engineering Center, 286pp.

US Army Corps of Engineers. 1997. *HEC-RAS River Analysis System User's Manual, Version 2.0*, Davis, CA: Hydrologic Engineering Center, 220pp.

Wiberg, P.L. and Smith, J.D. 1987. Calculations of critical shear stress for motion of uniform and heterogeneous sediments. *Water Resources Research* 23: 1471–1480.

Wilcock, P.R. 1997. The components of fractional transport rate. *Water Resources Research* 33: 247–258.

Wilcock, P.R. 2001. The flow, the bed and the transport interaction. In: Mosley, M.P., ed., *Gravel-bed Rivers V*, Christchurch: New Zealand Hydrological Society, pp. 183–220.

CHAPTER 19
Modelling fluvial morphodynamics

James E. Pizzuto
University of Delaware, Newark, DE, USA

19.1 Introduction

Fluvial morphodynamic models predict changes to the shape and character of fluvial landscapes through time. This chapter emphasizes quantitative rather than conceptual models, so morphodynamic modelling is viewed as a branch of applied mathematics. Rivers are influenced by mutually interrelated exogenous variables such as climate, vegetation, geological setting, tectonics and the activities of humans and other animals (beavers, for example), in addition to processes associated with flowing water. Morphodynamic modelling of rivers is a complex, relatively immature discipline that has grown dramatically in recent decades. In this chapter, we discuss examples of morphodynamic models where recent progress has been noteworthy. Some of the models provide interesting scientific insights, whereas others may be useful in river management and engineering.

The structure of a morphodynamic model includes defining the initial state of a fluvial landscape and determining its subsequent evolution through time (Fig. 19.1). Because rivers by definition involve water, morphodynamic models of rivers necessarily are coupled to those of the hydrologic cycle to specify the river's water discharge and hydraulic models to route flow through the river network.

The use of hydraulic models, however, introduces a fundamental conundrum of morphodynamic modelling: the rate of flow in the river is intimately coupled to the river's morphology, so there is a profound feedback between the processes that create a river's form and the form itself. For example, planar alluvial riverbeds are unstable when sediment is in motion (Parker 1976; Seminara 2010), and as a result they deform into characteristic landforms that include (at smaller scales) patches, streaks, ripples, dunes and antidunes, and (at larger scales) a variety of different types of bars (for definitions, see Knighton 1998). These landforms are interesting to study in themselves, but because they extract momentum from the flow and influence rates and patterns of sediment flux, they also influence larger scale fluvial landforms, including the channel's width, depth, slope and planform. As a result, morphodynamic models of larger features must necessarily account for the effects of these smaller features: all scales of fluvial morphology interact with each other and they are interdependent.

Hydraulic models themselves are insufficient, however, because changes in fluvial morphology are accomplished by erosion and deposition and therefore the hydraulic models must be coupled with models for the erosion of sediment and bedrock and the transport and deposition of sediment. Finally, the movement of sediment into and out of a reach must be related to morphological changes and all these computations must be extended through time. Additional exogenous models are needed to specify the supply of sediment, the growth of vegetation, changes in the geological setting, the activity of humans and other animals and so on. For example, humans will act to control and modify natural changes to an evolving river or the river will erode into bedrock and modify its geological setting and so on.

The process of morphodynamic modelling

Obtaining useful results from a predictive morphodynamic model requires a systematic approach that follows a series of important steps. These include [adapted from the protocol of Anderson and Woessner (1991) for groundwater modelling]:

1. *Establish the purpose* of the modelling study. Without a clearly defined purpose, the study cannot succeed.
2. Use field observations to *develop a conceptual model* of the problem to be solved. This is particularly important for modelling fluvial morphodynamics, as rivers are influenced by many different processes and all of them cannot be modelled simultaneously. Field data are needed to separate key processes from those that can be neglected.
3. The conceptual model is then rendered into *mathematical form* that can be solved numerically using a computer program.
4. *Model design* refers to the implementation of the computer program for a specific case. It involves developing a computational grid and selecting boundary conditions, relevant numerical values for parameters and other data that may be characteristic of a particular site or program. At this stage, detailed field data are always required for a site-specific modelling study.
5. *Calibration* involves adjusting the model design to reproduce a set of field observations. This is a standard practice in modelling and, though problematic, is nearly always required

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

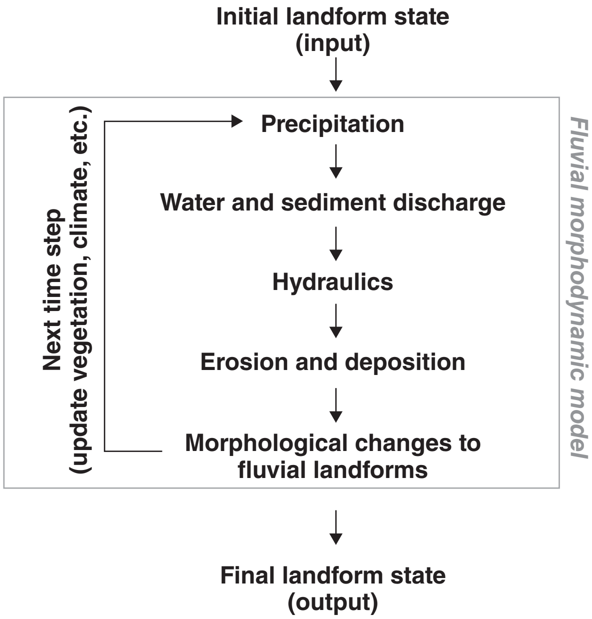


Figure 19.1 Components and structure of a fluvial morphodynamic model.

to demonstrate that the model is accurate enough to solve the problem that has been posed.

6 *Verification* is the process of applying the model to an independent set of field observations not used for calibration. This step provides an additional demonstration of the model's accuracy.

7 *Prediction* involves specifying future conditions. This usually satisfies the basic purpose of the modelling study.

Anderson and Woessner (1991) also advocate the use of post-audits in predictive modelling. The goal of a post-audit is to compare the predictions made in step 7 with new field data. These data allow the assumptions of the model to be evaluated. If the model's performance is not adequate, then the entire modelling process may need to be repeated, with appropriate corrections made at each step.

Morphodynamic modelling: science or art?

From this brief overview, it should not be surprising that fluvial morphodynamic modelling is often uncertain and perhaps more of an art than a precise science. Rivers evolve over long time-scales and defining correct initial conditions often requires reconstructing past fluvial landscapes from incomplete geological records. The relevant external controls of past climates or geological settings are equally uncertain. Even when such information can be obtained, our conceptual understanding of many fluvial processes remains poorly developed and a well-founded conceptual model of system behaviour is an essential foundation for model development. Finally, a strong conceptual understanding does not ensure that the relevant processes can be represented quantitatively, so predictive results may remain unattainable.

Two categories of fluvial morphodynamic models

Fluvial morphodynamic models may be classified in a variety of different ways, but it is most important to distinguish between two particular classes of rivers and the models that represent them. *Alluvial rivers* flow in valleys filled with sediment transported and deposited by fluvial action. These deposits are referred to as *alluvium*. The channels of alluvial rivers are carved into alluvium and changes in the morphology of alluvial rivers occur primarily through the erosion and deposition of alluvial sediment. *Bedrock rivers* have channels that are carved into bedrock. Changes in the morphology of bedrock rivers primarily occur through erosion of bedrock. Because sediment transport, erosion and deposition are very different processes from bedrock erosion, alluvial and bedrock rivers are modelled using different methods.

19.2 Modelling longitudinal profiles

The longitudinal profile of a river is defined by the stream's bed elevation as a function of distance downstream measured along the channel. Changes in the longitudinal profile through time can be accomplished though erosion of the bed material, deposition on the bed or tectonic activity (including both uplift and subsidence). Morphodynamic modelling of longitudinal profiles has provided interesting scientific results about the nature of fluvial systems and it is also been very useful in river management (particularly for alluvial rivers).

Alluvial rivers

Changes in the elevation of an alluvial streambed, z_0, through time (t), can be related to the supply rate of sediment entering and leaving an element of the bed, in addition to the rate of tectonic uplift, U. An expanded 'Exner' equation (Paola 2000) describes these principles:

$$\frac{\partial z_0}{\partial t} = -\frac{1}{1-p}\frac{\partial q_s}{\partial x} + U \qquad (19.1)$$

where p is the fractional porosity of the bed sediment, q_s is the volumetric bed material flux per unit width of channel ($m^3\ m^{-1}\ s$), x is the distance downstream (m) and U is the tectonic uplift rate (negative in the case of tectonic subsidence) ($m\ s^{-1}$). For most applications, the time-scale of tectonic processes is much slower than those associated with sediment transport and U can be neglected.

The bed material transport rate should depend on hydraulic processes, the morphology of the streambed and the nature of the bed material, so additional equations are needed to solve eqn. 19.1. To illustrate the nature of the problem, the simplest version of these could be written as

$$q_s = f\left(\frac{\partial z_0}{\partial x},\ \text{other terms}\right) \qquad (19.2)$$

Equation 19.2 indicates that the bed material flux depends on the local bed slope $\partial z_0/\partial x$ in addition to 'other terms'. Substituting eqn. 19.2 into eqn. 19.1:

$$\frac{\partial z_0}{\partial t} = -\frac{1}{1-p}\frac{\partial}{\partial x}\left[f\left(\frac{\partial z_0}{\partial x},\ \text{other terms}\right)\right] + U \qquad (19.3)$$

Equation 19.3 is a second-order differential equation with the bed elevation as the unknown variable. When combined with appropriate initial conditions (specifying the elevation of the bed at some initial time when computations begin) and boundary conditions, the solution to eqn. 19.3 provides the elevation of the streambed through time – a computational model for the development of a stream's longitudinal profile. The 'other terms' represent additional processes controlling forces exerted by fluids on the streambed sediment, including the development of ripples, dunes, bars and other small-scale bed features.

In gravel-bed rivers, the grain size of the bed surface is typically coarser than underlying sediment, a phenomenon referred to as 'pavement' or 'mobile armour'. The development of pavement represents a very strong control on fluvial processes and therefore it must be included in morphodynamic models of longitudinal profile evolution. This requires (i) specifying sediment transport processes for each grain size, (ii) accounting for how different grain sizes interact with the river bed and (iii) developing models to account for changes in grain size through time via grain abrasion, selective deposition and erosion and so on. Fortunately, theories for all these processes are well developed (for a thorough review, see Parker 2008).

Morphodynamic models of fluvial longitudinal profiles and bed surface texture have provided interesting insights to a variety of problems in recent decades. Sinha and Parker (1996) explain how subsidence, downstream fining by abrasion and wave-like aggradation can contribute to the development of longitudinal profile convexity, and Snow and Slingerland (1987) note that disequilibrium profiles rapidly evolve to resemble equilibrium forms, even when significant erosion remains before true equilibrium form is achieved. Paola (2000) and Robinson and Slingerland (1998) describe one-dimensional fluvial models that predict the distribution of fluvial deposits in subsiding sedimentary basins. Similar models also provide insights into the formation of ancient gold placer deposits (Slingerland *et al.* 1994). Theories to predict downstream fining in gravel-bed streams through aggradation and abrasion have been reasonably successful (Hoey and Ferguson 1994; Cui *et al.* 1996) and these models have also been used to explain the occurrence of rapid transitions between gravel-bed and sand-bed streams (Sambrook Smith and Ferguson 1995; Cui and Parker 1998; Parker and Cui 1998). Longitudinal profile models also provide useful insights into the erosion and downstream transport of episodic sediment inputs from landslides, mining and periodic storms (Lisle *et al.* 1997; Sutherland *et al.* 2002; Cui *et al.* 2003).

Longitudinal profile models have also been extensively used to predict patterns of erosion and deposition induced by a variety of watershed management activities. A thorough discussion of these applications is provided by Thomas and Chang (2008). It is important to remember that these models cannot resolve small-scale bed features such as pools and riffles or bars and therefore that all predictions will necessarily represent values averaged over channel lengths exceeding many channel widths (Cui *et al.* 2008). Additionally, lateral variations in transport processes cannot be explicitly represented, but must somehow be accounted for because channel width varies and transport processes do not extend across the entire wetted perimeter of a cross-section. Ferguson and Church (2009) advocate the use of an 'effective width' to account for complex lateral variations in bed material transport rates and they obtained useful predictions of transport and aggradation in their study of decadal evolution of the Fraser River in British Columbia, Canada, with this approach (Fig. 19.2)

Bedrock rivers

The longitudinal profiles of bedrock rivers evolve through bedrock incision and uplift, so the profile evolution equation, eqn. 19.1, can be written as

$$\frac{\partial z_0}{\partial t} = -E + U \qquad (19.4)$$

where E is the bedrock incision rate. The bedrock incision rate is typically represented as an empirical function of local river slope, drainage basin area (a surrogate for water discharge) and lithology (Whipple *et al.* 2000).

Rates of bedrock incision are typically much lower than rates of erosion and deposition in alluvial rivers, so morphodynamic models of bedrock rivers have not been widely used in river management. During the last few decades, however, bedrock river models have led to important insights into the relative impacts of tectonics, climate (Roe *et al.* 2002) and lithology (Stock and Montgomery 1999) on the evolution of fluvial longitudinal profiles developed on bedrock surfaces (Whipple 2004).

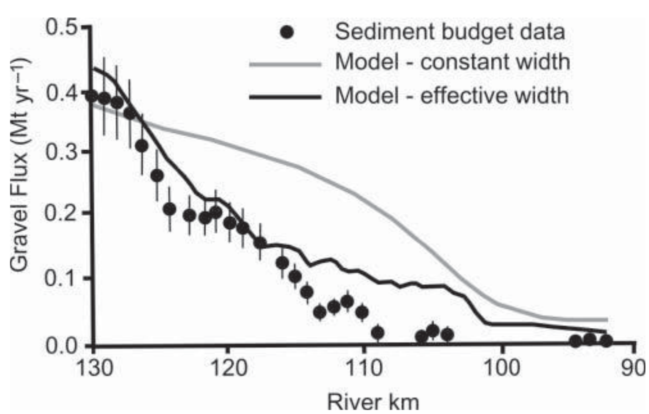


Figure 19.2 Gravel bedload transport rates determined from a sediment budget and computed using a one-dimensional sediment transport model for a section of the Fraser River in British Columbia, Canada. Predicted transport rates using models with constant width and spatially variable effective width are indicated. Adapted from Ferguson *et al*, 2009.

Research on bedrock rivers is proceeding with great intensity. New erosion models are being developed to account better for the influence of erosive sediment 'tools' (Sklar and Dietrich 2004; Turowski *et al.* 2007), to include specific erosional processes such as abrasion and plucking (Chatanantavet and Parker 2009), to account for bedrock weathering processes (Hancock *et al.* 2011) and to understand better when bedrock erosion occurs through debris flows rather than traditional fluvial processes (Stock and Dietrich 2006). Mixed bedrock–alluvial channels are also increasingly being described (Turowski *et al.* 2008; Skalak and Pizzuto 2010;Nittrouer *et al.* 2011) and new models are being developed for these hybrid bedrock-alluvial systems (Pizzuto *et al.* 2008; Nelson and Seminara 2012).

19.3 Modelling hydraulic geometry of rivers

Geomorphologists (Davis 1899; Mackin 1948) and engineers (Blench 1969) have long hypothesized that rivers and canals attempt to create a form that is in equilibrium with prevailing discharge, sediment supply and other constraints. Leopold and Maddock (1953) noted that, for many rivers, width, depth, slope and velocity, when plotted against discharge of a constant recurrence interval in a watershed, define power functions with nearly universal exponents. Leopold and Maddock (1953) referred to these functions as hydraulic geometry equations and they have been widely interpreted to represent evidence for the existence of a quasi-equilibrium state (Wolman 1955), although the hydraulic geometry equations appear to be satisfied even when fluvial systems are changing rapidly (Knox 1976).

The ability to understand and predict the equilibrium morphology (width, depth, slope and planform) of a river is of great practical importance. It is obviously essential for designing channels that are able to maintain their form without excessive erosion or deposition, criteria that are paramount for many river engineering and restoration projects. Understanding changes in hydraulic geometry, and the time-scales over which they occur, have proven helpful in assessing the effects of urbanization (Hammer 1973; Leopold 1973), riparian vegetation (Andrews 1984; Hey and Thorne 1986) and other influences on river channel form and process.

The discovery of empirical hydraulic geometry equations stimulated many efforts to explain the principles that underlie them, to provide theoretical explanations for a river's width, depth and slope given the supply and calibre of sediment, the water discharge provided by the drainage basin and other relevant variables. Some of the basic concepts of these theories are briefly outlined here. More thorough discussions are available elsewhere (Parker 2008; Pizzuto *et al.* 2008; Buffington 2012).

Hydraulic geometry theories can be divided into two types: one type considers the physical processes that must exist to ensure channel stability, whereas the other uses optimization criteria to determine hydraulic geometry.

For a given channel slope, a stable alluvial channel cross-section can exist if the water flowing downstream carries no sediment. This observation, although correct, does not lead to a unique solution for the channel geometry, because if any channel is lined with large particles that cannot be moved and if no sediment is carried into the reach from upstream, the channel form must continue to exist without changing its form.

The idea of a channel lined with immovable sediment provided the first physically based hydraulic geometry theory. Lane (1957) suggested that channels could exist where sediment along the entire perimeter is precisely at the threshold of sediment motion: poised to move (at the ambient steady water discharge), yet remaining in place. For non-cohesive sediment with a uniform grain size, this hypothesis leads to a solution for the channel cross-sectional shape and size given an imposed slope and a steady, channel-forming discharge (Fig. 19.3a) (Lane 1957; Diplas and Vigilar 1992).

It is not clear how the concept of a threshold channel can be applied to natural alluvial channels that form in alluvium. For the alluvium to be stored in a river valley, sediment must

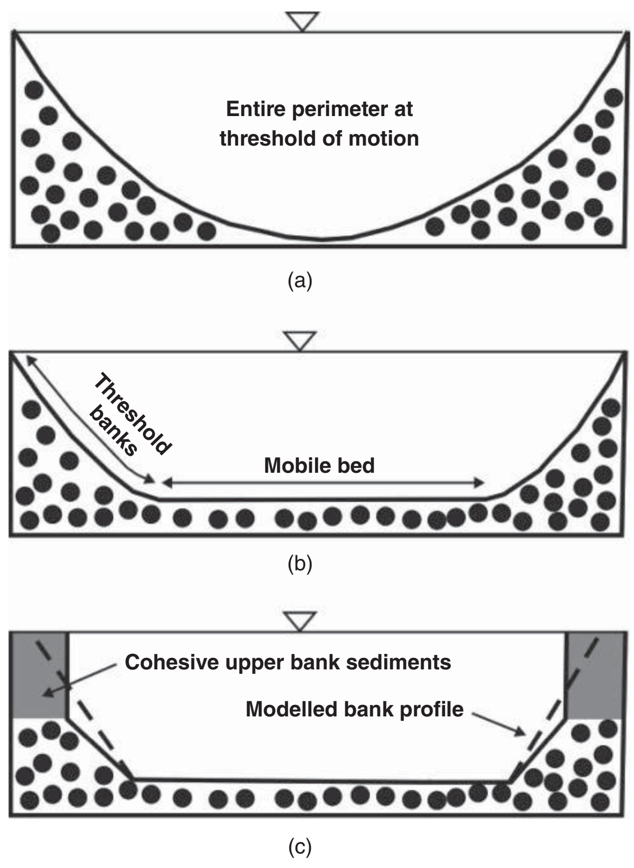


Figure 19.3 Channel cross-sectional morphology predicted by different hydraulic geometry models. (a) Threshold channel in non-cohesive sediment. After Diplas and Vigilar (1992). (b) Gravel bed river carved into non-cohesive sediment with threshold banks and active bed. After Parker (1978b, 1979). (c) Composite bank with overlying cohesive overbank deposits and trapezoidal channel representation. Adapted from Diplas, 1992.

be transported in place and deposited, yet threshold channels transport no sediment. The concept might apply to steep mountainous terrain where sediment is supplied by hillslope processes or debris flows and subsequently slowly moved grain-by-grain until a stable channel is formed that no longer evolves, but clearly most alluvial river channels cannot be threshold channels (although the concept might be useful for engineering design).

If the threshold channel theory could be extended to admit the possibility of sediment movement along the bed of the stream, but under the additional constraint that the banks remain stable, without net erosion or deposition, then such a channel could be stable and still carry bed material downstream. If the bank morphology is adjusted so that the sediment is precisely at the threshold of sediment motion, then the bank morphology and the channel morphology could be specified.

This problem was originally posed and solved by Parker (1978b, 1979), whose solution was extended by Vigilar and Diplas (1997) (Fig. 19.3b). These theories only apply to gravel-bed streams without bank vegetation: stable banks cannot be maintained at the threshold of motion in finer grained sediment (sand, for example). The theories predict that the depth of gravel-bed channels will be about 20% greater than those of threshold channels for a given slope and grain size, yielding a constant Shields stress (Knighton 1998; Garcia 2007) on the channel bed that is about 20% greater than the threshold value required to mobilize the bed sediment. This prediction is reasonably well satisfied by many gravel-bed streams (Andrews 1984; Parker *et al.* 2007) (Fig. 19.4), and the initial theory has been extended to include variable grain size (Ikeda *et al.* 1988) and rigid bank vegetation (Ikeda and Izumi 1990).

For rivers that transport sediment in suspension, deposition should occur somewhere across the channel perimeter, most likely along the banks where the flow is relatively quiescent. For an equilibrium cross-section to exist, this deposition must be balanced by an equivalent amount of erosion. Parker (1978a) suggested that the cross-section of a stable suspended load stream should be adjusted so that erosion and deposition on the banks should everywhere be equal, and he used this observation to derive hydraulic geometry equations for sand-bed streams. Although Parker's (1978a) hypothesis seems reasonable, this approach has not been verified by field observations.

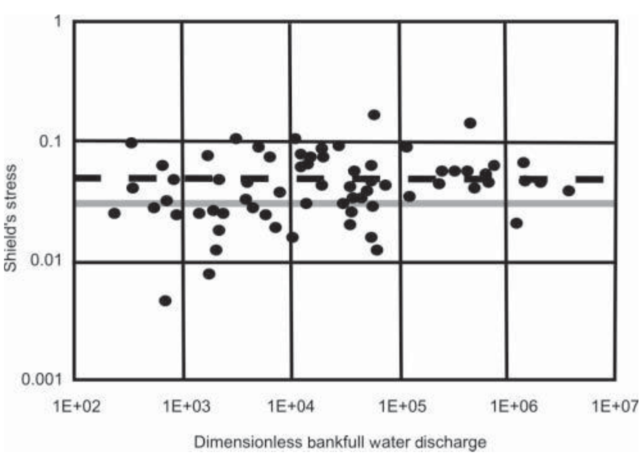


Figure 19.4 Bankfull Shields stress (solid circles) as a function of dimensionless water discharge for selected gravel-bed rivers. Critical Shield's stress required to move bed material (solid grey line, Shield's stress = 0.03) and average bankfull Shield's stress for the data (black dashed line, Shield's stress = 0.0489) are also indicated. Redrafted from Parker *et al.* (2007), where variables are explicitly defined.

Perhaps because of the difficulty of testing physically based models for stable hydraulic geometry, a wealth of optimization hypotheses have been proposed. These hypotheses suggest that rivers adjust their morphology to transport the imposed water and sediment discharge, subject to an additional constraint that optimizes another hydraulic process. Specific optimization hypotheses vary widely. A review by Eaton and Church (2007) summarizes the optimization principles used by different authors, stating that they include 'minimum variance of hydraulic quantities (Langbein and Leopold 1966) …, minimum unit stream power, maximum sediment transport efficiency (Kirkby 1977), minimum power expenditure (Chang 1979), maximum sediment-transporting capacity (White *et al.* 1982), least action (Huang and Nanson 2000) or minimum energy expenditure (Huang *et al.* 2004)'. Eaton and Church (2007) proposed another criterion, suggesting that the 'stable state is the one in which resistance to flow is maximized subject to the condition that the imposed water and sediment load – that is, the load delivered from upstream – be passed on the adopted channel gradient … this condition is mathematically equivalent to the condition that the imposed sediment load be passed with the smallest expenditure of energy'.

Despite the importance of the problem and decades of focused research, a widely accepted solution for the hydraulic geometry problem remains elusive. Geomorphologists and hydraulic engineers even fail to agree on a conceptual model for how channels create a stable cross-section and, as a result, both physically based (Parker *et al.* 2007) and optimization approaches (Eaton and Church 2007) continue to be proposed for alluvial river channels (e.g. Fig. 19.3c). Remarkably, both methods appear to provide useful and compelling results, despite the fundamental differences between them.

It is relatively straightforward, however, to propose a conceptual framework for solving the hydraulic geometry problem: a stable channel will exist when the rates of erosion and deposition, averaged over suitable spatial and temporal scales, will be in balance. Unfortunately, quantifying this principle requires detailed models for a variety of processes that are poorly understood. For example, rates of erosion and deposition on vegetated, cohesive banks must be quantified over decadal time-scales that include widely varying flows (because erosion tends to occur at high flows, whereas deposition is favoured at lower flows) (Pizzuto 1994). Models for bank erosion continue to evolve and improve, but models for bank deposition remain poorly developed (Pizzuto and the ASCE Task Committee 2008). Solving the hydraulic geometry problem will also involve understanding floodplain evolution, because the bankfull depth is set by the difference in elevation between the ambient floodplain and the streambed. Furthermore, the channel slope is set by the river's planform (meandering reduces a river's slope, for example), so a physically based model for fluvial hydraulic geometry must include a model for channel planform evolution. Assuring that a channel can transport the imposed load of water and sediment requires accurate models of sediment transport, frictional resistance to flow and so on.

Solving for a river's reach-averaged width, depth and slope appears to be a relatively straightforward problem. From a mathematical point of view, it only requires solving three equations. However, a river's hydraulic geometry is determined by all components of the entire fluvial system operating through time. Therefore, although the solution is easy to pose (a reach-averaged balance between erosion and deposition), quantifying it requires a profound understanding of fluvial processes that continues to elude geomorphologists.

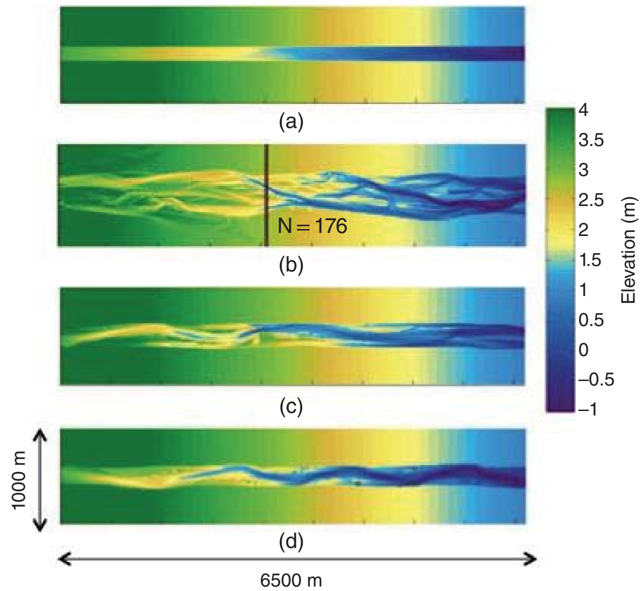


Figure 19.5 Numerical model results of river planforms development over 10 years with different floodplain vegetation. (a) Initial configuration; (b) without floodplain vegetation (note development of braided channel planform); (c) with low-density 'pioneer' vegetation on floodplain (note sinuous channel planform); (d) with grass (note confined sinuous channel). Source: Crosato and Saleh, 2011. Reproduced with permission of Wiley.

19.4 Modelling channel planforms

Rivers viewed from above the Earth's surface present a fascinating variety of forms, commonly referred to by geomorphologists as river planforms. Studies of river planforms have generally focused on alluvial rivers, despite the widespread influence of bedrock on river planform geometry, and as a result here we focus exclusively on morphodynamic models of alluvial river channel planforms. Morphodynamic models of bedrock channel planforms are poorly developed, but interest is growing (e.g. Finnegan and Dietrich 2011). Following two decades of progress in modelling the evolution of bedrock river profiles and growing interest in mechanisms of fluvial bedrock erosion, rapid future progress in modelling bedrock river planforms is likely.

One fundamental attribute of alluvial rivers is whether they have a single channel ('single-thread') or multiple channels ('branched', 'anastomosing' or 'braided'). The individual channels of anastomosing rivers can display all the forms that are typical of single-thread rivers. Kleinhans (2010) presented a useful review of our understanding of alluvial channel planforms.

The goal of morphodynamic modelling of river channel planforms is to predict the development of different channel planforms within a valley filled with alluvium. Successful models must utilize the complete spectrum of discharges that occur in nature and the growth of vegetation and its morphological impacts will necessarily be included. Humans are increasingly viewed as dominant geomorphic agents (Hooke 1994; Merritts *et al.* 2011) and successful models of river channel planforms will also have to include anthropogenic impacts.

Current research falls far short of this goal, although progress in recent decades has been truly remarkable. Most simulations that attempt to predict river planforms within an entire alluvial valley are in their infancy, with detailed simulations only capable of representing short valley segments over a few decades of morphodynamic evolution using two- and three-dimensional hydrodynamic models (Fig. 19.5) (Crosato and Saleh 2011, Li and Millar 2011). Reduced complexity models (Coulthard and Van De Wiel 2006, 2012) can simulate processes over longer spatial and temporal scales, but these models lack spatial resolution and physical processes are highly generalized.

Rather than predicting the occurrence of different channel planforms, current models are more successful when predicting how channels evolve once planforms are specified. Models for meandering channels are the most highly developed, whereas models of braided and anastomosing channels remain in their infancy.

Meandering channels

Mathematical theories for the origin of meandering channels are so well developed that this field can be described as 'mature'. According to theory, meandering channels can develop from straight channels as a result of two hydrodynamic instabilities, one associated with sediment transport over a granular stream bed and the other related to channel curvature. In relatively narrow channels, the former instability creates regularly spaced, rapidly migrating alternate bars. The alternate bars deflect the flow into the banks, leading to bank erosion and incipient channel curvature. At certain bend wavelengths, growing bends create 'forced' bars that do not migrate and the channel bed and planform gradually evolve into the characteristic morphology observed in meandering streams. Classic studies of these processes include those by Hasegawa (1977), Ikeda *et al.* (1981), Howard and Knutson (1984) and Johanneson and Parker (1989), with more recent work summarized by Seminara (2006, 2010).

Theories of meandering require quantitative expressions for the flow in curving channels, models for bed material transport (suspended load is typically ignored) and the evolution of the bed topography and a means of predicting lateral channel migration. Hydrodynamics are typically expressed in terms of a quasi-one-dimensional model that predicts the flow field based on the curvature of the channel centreline and the reach-averaged channel morphology (width, depth and slope, all assumed to be constant). Bed material transport and bed topography are predicted using standard expressions (Seminara 2006; Garcia 2007). Typically, lateral channel shifting has been computed using an equation such as

$$V = E(u - U) \tag{19.5}$$

Where

- V is the rate of bank retreat in a direction perpendicular to the channel centreline,
- U is the reach-averaged velocity,
- u is the velocity 'near' the 'outer' bank of a meander bend and
- E is a dimensionless constant that reflects the erodibility of the bank materials.

Deposition on the inner bank has not been treated explicitly in most analyses and the width is typically assumed to be constant. Thus, when bank erosion occurs, the channel is simply shifted laterally.

Mathematical theories of meandering have fostered the development of numerical models applied to both scientific and applied problems. Ikeda *et al.* (1981) showed that the theory can explain the development of broadly looping meander bends they termed 'Kinoshita' curves and Furbish (1991) used the theory to describe the variability of meander morphology and spatial patterns in channel migration rates. Meander migration theories do not include processes of flow separation on the insides of tight bends, but the theories can explain observed relationships between channel migration rate and bend curvature (Furbish 1988), suggesting that flow separation does not strongly influence spatial patterns in lateral migration rates (Crosato 2009). Stolum (1996) argues that meanders self-organize into a critical state characterized by fractal geometry, whereas Frascati and Lanzoni (2010) used a fully non-linear simulation to suggest that long-term meandering dynamics are not consistent with the presence of chaos and self-organized criticality. Howard (1992, 1996), Sun *et al.* (1996), Gross and Small (1998) and Xu *et al.* (2011) coupled meander migration models with schemes for floodplain deposition, allowing interaction of meanders with evolving floodplain deposits (Fig. 19.6). These studies demonstrate that meander morphology is highly influenced by variations in floodplain morphology and erodibility, a conclusion reinforced by theoretical and empirical studies of Guneralp and Rhoads (2011). Larsen and Greco (2002) used Parker's meander migration model to evaluate the long-term impacts of bank stabilization designs. Constantine *et al.* (2009) demonstrated that the bank erodibility coefficient E can be predicted using a jet-test device that measures soil erosion resistance to fluid shear stresses.

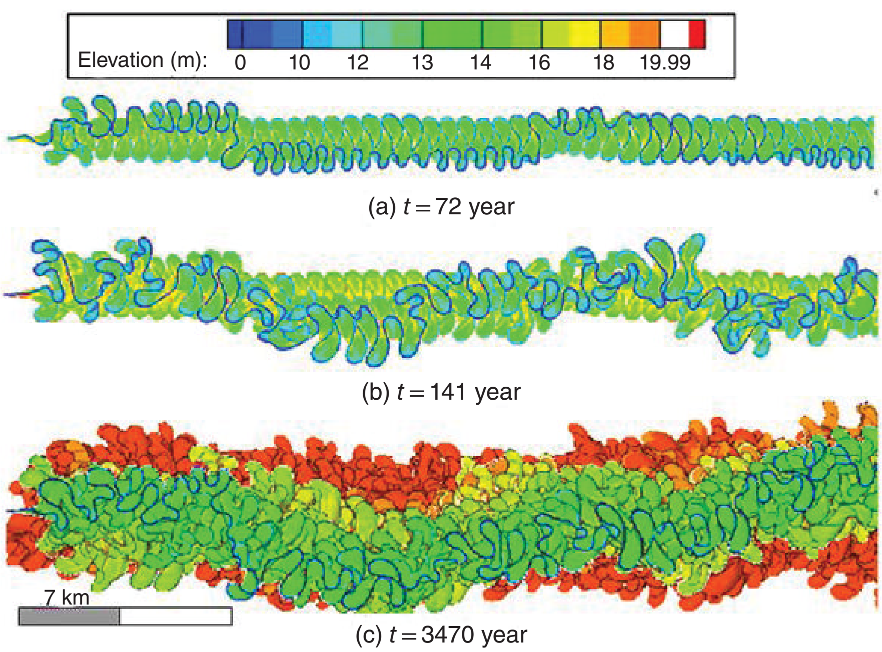


Figure 19.6 History of floodplain deposition predicted by a numerical model of river meandering over a period of 3470 years. Source: Xu *et al*, 2011. Reproduced with permission of Elsevier.

Recent research is rapidly extending the classic mathematical theory of meandering. Bank erosion processes have received some attention: Dulal at al. (2010) and Parker *et al.* (2011) describe how cohesive floodplain deposits can retard lateral migration rates, while Xu *et al.* (2011) explicitly include the influence of varying bank height on meander migration rates (they also model floodplain accretion). The width of meandering rivers is not constant, but varies systematically (Luchi *et al.* 2011) (Fig. 19.7). Luchi *et al.* (2012) present a theory to explain why width is a maximum at bend apices in some rivers, but a minimum at bend apices in other rivers. Parker *et al.* (2011) present a model that explicitly treats the advance and retreat of both river banks independently, allowing the width to vary both temporally and spatially in response to hydraulic and sediment transport processes. Many researchers are exploring the utility of two- and three-dimensional hydraulic models to predict the evolution of meandering stream channels. These models allow (and require) increasingly complex algorithms for bank erosion and retreat (Duan and Julien 2010), but better capture interactions between complex bed topography, bank morphology and river channel form. Two- and three-dimensional computational models of meandering are currently limited by the extensive computational resources required to simulate the development of long meander trains that evolve over geological time-scales.

Braided channels

The first numerical model to simulate successfully the formation of braided channels was the rule-based cellular automaton scheme of Murray and Paola (1994). Their model is based on a series of rules for routing sediment and water between the rectangular cells that comprise the model domain. Water is routed into downstream cells according to the local bed slope and sediment is routed according to idealized sediment transport equations. Murray and Paola (1994) concluded from their simulations that braiding develops when bedload transport leads to excess scour in flow convergences and deposition in divergences. This in turn requires that the flow be sufficiently unconstrained laterally, that it can change its width freely and, if discharge or stream power is used to parameterize sediment flux, that the exponent in the sediment-flux law be >1. The model results suggest that braiding is a simple, robust result of bedload transport under these conditions. The ubiquity of braiding in model runs suggests that braiding may be *the* fundamental instability of laterally unconstrained free-surface flow over cohesionless beds. This conclusion is supported by hydrodynamic stability analyses, which indicate that sediment transport over a plane bed is fundamentally unstable, leading to the creation of mid-channel bars (and, ultimately, braided channels) in wide rivers (Parker 1976; Seminara 2010).

Murray and Paola's (1994) initial approach has been extended considerably. Murray and Paola (2003) added 'rules' to represent vegetation-induced cohesion and their results indicate that vegetation promotes the development of a single-thread channel rather than a braided channel. Coulthard *et al.*'s cellular model, CAESAR, uses an improved routing scheme for water and sediment (Coulthard *et al.* 2002). In the last decade, CAESAR has been used to study the development of alluvial fans, storage and reworking of metals in fluvial environments, the impacts of changing climate, vegetation and land use on fluvial landscapes and other interesting questions, reviewed by Coulthard *et al.* (2007) and Coulthard and Van De Wiel (2012). Modifications allow CAESAR to simulate some of the processes found in meandering channels (Van De Wiel *et al.* 2007).

Although the simplicity of cellular models such as CAESAR allows long time-scales and large spatial scales to be simulated, many hydraulic and sediment transport processes remain highly idealized, limiting their utility. Some recent studies have shown promising simulations of braided channel processes (Kleinhans *et al.* 2008; Crosato and Saleh 2011) using two- and three-dimensional hydrodynamic models (Fig. 19.8), but these methods require enormous computational resources that limit their utility at present.

Anastomosing channels

Morphodynamic modelling of anastomosing channels has attracted relatively little attention, although recent advances in

Figure 19.7 Meander bends showing systematic variations in width. (a) Maximum widths at bend apices on the Brazos River in Texas and (b) minimum widths at bend apices and widening in straight reaches along a tributary of the Amazon River in Brazil. Source: Luchi *et al*, 2012. Reproduced with permission of Google Maps.

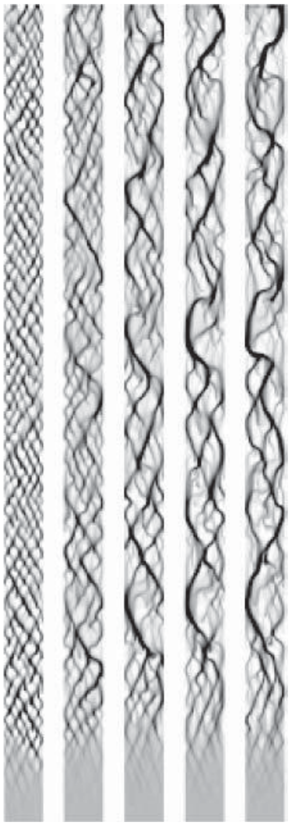


Figure 19.8 Evolution of a braided channel planform computed using Delft3D, loosely based on a channel 2000 m wide and 40 km long similar to the river Rhine. Panels illustrate (from left to right) planform development after 30, 50, 70, 90 and 110 years. See Kleinhans (2010) for additional details.

the modelling of delta morphodynamics suggest that future progress may be rapid. The complex processes that create stable, multiple-thread rivers resist many of the methods used to analyse the evolution of single-thread channels, so initial efforts have relied on reduced complexity, rule-based modelling. Murray and Paola's (2003) cellular model that incorporates vegetation-induced cohesion creates channels that appear to be similar to anastomosing channels, although the authors refer to them as 'multiple channel (braided) patterns' rather than characterizing them as anastomosing. Anastomosing channels are created by models of alluvial channel architecture that incorporate avulsion (e.g. Karssenberg and Bridge 2008); these models generate anastomosing channels using stochastic 'rules' for avulsion based on the cross-valley and down-valley slope of the floodplain and the annual flood discharge.

Rather than trying to develop models that create entire anastomosing systems, recent research is in an initial phase that focuses on the morphology and stability of individual channel bifurcations. Progress has been made in determining the division of water and sediment in divided channels (Kleinhans *et al.* 2008), explaining why stable bifurcations are typically of unequal sizes (Edmonds and Slingerland 2008) and documenting the influence of upstream meanders (Kleinhans *et al.* 2008, 2011), width adjustment (Kleinhans *et al.* 2011) and bar dynamics (Kleinhans *et al.* 2008) on the evolution single bifurcations. Edmonds and Slingerland (2010) showed that sediment cohesion strongly influences the morphology of divided channel systems.

19.5 Modelling floodplain sedimentation and erosion

Introduction

Geomorphic processes on floodplains are remarkably diverse. Some floodplains are dominated by lateral channel migration and therefore models of in-channel processes are central to predicting their evolution. Others may be dominated by sedimentation and erosion during overbank flows. Many floodplains are strongly influenced by vegetation. On still others, episodic rare storms may completely obliterate floodplain deposits (Nanson 1986), while in other settings floodplains are barely affected by rare, high-magnitude storms (Costa 1974). Many floodplains have been dramatically affected by recent human activities and these cannot be considered 'quasi-equilibrium' geomorphic systems (Merrits *et al.* 2011). In their classification of floodplains, Nanson and Croke (1992) described an incredible variety of different floodplain types, each of which reflects different characteristic processes.

Well-developed conceptual models of floodplain sedimentation and evolution are limited to a few end-member cases. Conceptual models of floodplains along meandering rivers are most highly developed. Conceptual models exist for floodplains along braided rivers and some types of anastomosing channels, but these are considerably less well developed. Evolutionary trajectories for other types of floodplains are virtually unknown except through isolated case studies. Quantitative models cannot be developed without a strong conceptual understanding and, as a result, morphodynamic models for floodplain systems are best developed for meandering streams and are in their infancy for all other types.

As is typical for any geomorphic system, temporal and spatial scales dictate the processes that must be considered in floodplain morphodynamic modelling. Over very short time-scales, geomorphic setting (i.e. channel and floodplain geometry, vegetation characteristics, etc.) may be held constant and modelling floodplain geomorphic processes may be accomplished by detailed hydrodynamic and sediment transport models that provide considerable spatial and temporal resolution. Over longer time-scales, however, channel position, floodplain morphology, vegetation type and other characteristics will all vary, driven by temporal changes in climate, land use and the intrinsic operation of the fluvial system. Under these conditions, models must predict all these variables. Owing to limited computing resources and lack of thorough understanding, models become increasingly idealized and imprecise with increasing spatial and temporal scales. Unfortunately, time-scales for floodplain evolution are typically long (hundreds, thousands and even millions of years) and processes that govern floodplain evolution have large characteristic spatial scales. Floodplain morphodynamics modelling is particularly challenging and remains in its infancy.

Modelling event-scale floodplain processes

One-dimensional hydrodynamic models that can predict spatial and temporal patterns of flood inundation have been available for decades (French 1985). When coupled with algorithms for sediment transport and bank erosion, these models can route sediment downstream through valley systems over time-scales that represent multiple flood events (Carroll *et al.* 2004; Thomas and Chang 2007). Although these predictions are useful for many purposes, they provide little information on fluvial morphodynamics and are ultimately of limited use in routing sediment, because the relevant processes are difficult to capture in one dimension.

Recent improvements in numerical methods and greatly increased computational power have led to the development of two- and three-dimensional computational fluid dynamics models (Coulthard and Van De Wiel 2012) that can simulate floodplain processes (Nicholas *et al.* 2006, Thonon *et al.* 2007; Arboleda *et al.* 2010). These models typically simulate floodplain inundation and selected sedimentation processes over relatively short reaches of tens to hundreds of river widths and time-scales of years, decades and occasionally a century or so. Owing to these relatively short spatial and temporal scales, these studies typically focus on overbank sedimentation patterns, rather than the complete morphodynamic evolution of floodplains. However, rapid progress in the development and application of these models is likely in the near future.

Geomorphic models of floodplain evolution

The simplest models of floodplain morphodynamics combine meander migration models with simple models of floodplain accretion based on elevation (Howard 1996; Xu *et al.* 2011) or distance from an active channel (Gross and Small 1998). These methods create realistic-looking floodplain topography and they also reproduce commonly observed geomorphic features such as oxbow lakes and scroll-bars (Fig. 19.6). However, because they neglect spatially variable hydrodynamic processes during overbank flows, they provide little insight into the evolution of real floodplains.

The ability to simulate channel-scale processes within landscape evolution models has provided new tools that can simulate generalized floodplain evolution on watershed scales over significant geological time periods (reviewed by Coulthard and van de Wiel 2012; Baartman *et al.* 2012 provide a summary of the model LAPSUS). These models route water and sediments on a grid that can cover an entire watershed. Refinements in model conceptualization and spatial discretization have allowed channel and floodplain processes in watersheds to be simulated. Although initially these models were used for generalized theoretical purposes, they are now actively being used to simulate real landscapes (see Coulthard and van de Wiel 2012 for examples). The strength of these models lies in their ability to simulate processes over geological timescales and to include important coupling between exogenous variables (Fig. 19.1) and the fluvial system. They remain unable to include many detailed hydrodynamic and sedimentation processes, so although their capabilities show great promise, further research is needed to explore their potential fully.

The future of floodplain morphodynamic modelling

Floodplains play an important role in watershed management, as they store contaminated sediment, provide important ecological services, diminish flood peaks and in part control and reflect channel form and function. As a result, developing accurate morphodynamic models for floodplains should be a research priority.

Fortunately, rapid advances are likely in the near future. Sophisticated numerical models of floodplain hydrodynamics and sedimentation are increasingly available (as an example, DELFT3D can now be freely downloaded at http://oss.deltares.nl/web/delft3d). As noted above, our understanding of important floodplain processes is improving rapidly in certain areas. For example, the influence of vegetation on hydraulics and sedimentation is increasingly included in some of the new models.

Additional research will be needed, however. Our conceptual understanding of the evolution of many types of floodplains is inadequate to support the development of quantitative models. Models of the transport mechanics of fine sand, silt and clay-sized particles are poorly developed and these sediments are the primary components of many floodplains. Finally, modelling strategies must be developed that can be applied to large watersheds with an entire network of floodplains, possibly totalling thousands of miles in length. This will require distilling floodplain morphodynamic models into relatively simple algorithms that can account for floodplain sediment storage and remobilization. Compact analytical models may even be necessary (e.g. Pizzuto 2012).

19.6 Conclusion

The models reviewed in this chapter have several limitations in common (Table 19.1). With the possible exception of one-dimensional models of alluvial river bed profiles, few of the models are routinely used in practical applications. Few models are commercially available, all are difficult to use and few have been thoroughly tested in a wide variety of field settings. All models require field data for calibration, parameterization or initial and boundary conditions, and such data are difficult to obtain. Finally, it is important to recognize that the conceptual basis for many of the models discussed here is still being actively

Table 19.1 Limitations of numerical models discussed in this chapter.

Limitations	Possible solutions
Models are difficult to use	Develop user-friendly interface; only experts should use them
Conceptual models of system behaviour poorly developed	Detailed field work, further model development
Parameterization of sediment transport processes is imprecise	Monitoring studies to calibrate transport models
Long time-scales make model calibration imprecise or impossible	Calibrate model (at least partially) using stratigraphic data
Suitable boundary/initial conditions are unavailable	Detailed field work

debated. Scientists still do not entirely agree on the essential controlling mechanisms for many of the processes simulated by these models.

These observations have important implications for how such models should be used (Table 19.2). First, detailed field studies should always accompany any modelling study that is designed to predict the behaviour of any particular river. These field studies should not simply seek to provide parameter values needed for modelling, but should rather be comprehensive enough so that the modelling team can understand what processes control the river's behaviour and so that relevant historical influences on river channel form can be identified. Broadly trained fluvial geomorphologists are perhaps best suited for these types of investigations. Second, modelling should only be performed by those intimately familiar with the mathematical and physical basis for the models being used. General experience with numerical methods will not be sufficient for these models to be used properly, as all of the models are based on limited empirical data and simplifying assumptions that are often not fully appreciated or understood.

Despite these caveats, however, morphodynamic modelling of river systems is on the verge of an explosion of progress. Increasing computational efficiency and improved understanding of a host of fluvial processes (bedload transport rates of individual grain size fractions, bank erosion, vegetation effects and others) are catalysing the development of new models of river behaviour. New methods of collecting data over large areas using LiDAR and other remote sensing tools and improved physical modelling techniques provide new methods for calibrating and testing morphodynamic models. These models will be needed to achieve sustainable management of fluvial landscapes that are increasingly influenced by global climate change and human impacts.

Table 19.2 Ability of selected numerical models reviewed in this chapter to solve site-specific problems.

Subject	Relative utility	Comments
One-dimensional models of alluvial channel profiles	High	Calibration required, channel form must be constant
Quasi-equilibrium cross-sectional morphology	Poor	Conceptual basis poorly established
Meander evolution	High	Relatively untested
Braided channel evolution	Moderate	Available models promising but poorly tested and little used
Floodplain processes	Moderate	Rapid future progress likely

References

Anderson, M.P. and Woessner, W.W. 1991. *Applied Groundwater Modeling: Simulation of Flow and Advective Transport*, London: Academic Press, 381pp.

Andrews, E.D. 1984. Bed material entrainment and hydraulic geometry of gravel-bed rivers in Colorado. *Geological Society of America Bulletin* 95: 371–378.

Arboleda, A.M., Crosato, A. and Middelkoop, H. 2010. Reconstructing the early 19th-century Waal River by means of a 2D physics-based numerical model. *Hydrological Processes* 24: 3661–3675.

Baartman, J.E.M., van Gorp, W., Temme, A.J.A.M. and Schoorl, J.M. 2012. Modelling sediment dynamics due to hillslope–river interactions: incorporating fluvial behavior in landscape evolution model LAPSUS. *Earth Surface Processes and Landforms* 37: 923–935.

Blench, T. 1969. *Mobile-bed Fluviology: a Regime Treatment of Canals and Rivers*, Edmonton, AB: University of Alberta Press, 168pp.

Buffington, J.M. 2012. Changes in channel morphology over human time scales. In: Church, M., Biron, P.M. and Roy, A.G., eds., *Gravel-bed Rivers: Processes,* Tools, Environments, Chichester: John Wiley & Sons, pp. 435–463.

Carroll, R.W. James, A.,Warwick, J.J. and Miller, J. 2004. Modeling erosion and overbank deposition during extreme flood conditions on the Carson River, Nevada. *Journal of Hydrology* 297: 1–21.

Chang 1979. Minimum stream power and river channel patterns. *Journal of Hydrology* 41: 303–327.

Chatanantavet, P. and Parker, G. 2009. Physically based modeling of bedrock incision by abrasion, plucking and macroabrasion. *Journal of Geophysical Research* 114: F04018.

Constantine, C.R., Dunne, T. and Hanson, G.J. 2009. Examining the physical meaning of the bank erosion coefficient used in meander migration modeling. *Geomorphology* 106: 242–252.

Costa, J.E. 1974. Response and recovery of a piedmont watershed from tropical storm Agnes, June 1972. *Water Resources Research* 10: 106–112.

Coulthard, T.J. and Van De Wiel, M. J. 2006. A cellular model of river meandering. *Earth Surface Processes and Landforms* 31: 123–132.

Coulthard, T.J. and Van De Wiel, M.J. 2012. Modelling river history and evolution. *Philosophical Transactions of the Royal Society A* 370: 2123–2142.

Coulthard, T.J., Macklin, M.G. and Kirkby, M.J. 2002. A cellular model of Holocene upland river basin and alluvial fan evolution. *Earth Surface Processes and Landforms* 27: 269–288.

Coulthard, T.J., Hicks, D.M. and Van De Wiel, M.J. 2007. Cellular modelling of river catchments and reaches: advantages, limitations and prospects. *Geomorphology* 90: 192–207.

Crosato, A. 2009. Physical explanations of variations in river meander migration rates from model comparison. *Earth Surface Processes and Landforms* 34: 2078–2086.

Crosato, A. and Saleh, M.S. 2011. Numerical study on the effects of floodplain vegetation on river planform style. *Earth Surface Processes and Landforms* 36: 711–720.

Cui, Y. and Parker, G. 1998. The arrested gravel front: stable gravel-sand transitions in rivers. Part 2. General numerical solution. *Journal of Hydraulic Research* 36: 159–182.

Cui, Y., Parker, G. and Paola, C. 1996. Numerical simulation of aggradation and downstream fining. *Journal of Hydraulic Research* 34: 184–204.

Cui, Y., Parker, G., Pizzuto, J. and Lisle, T.E. 2003. Sediment pulses in mountain rivers. 2: Comparison between experiments and numerical predictions. Water Resources Research, 39(9): 1240.

Cui, Y., Wooster, J., Venditti, J., Dusterhoff, S., Dietrich, W. and Sklar, L. 2008. Simulating sediment transport in a flume with forced pool-riffle morphology: examinations of two one-dimensional numerical models. *Journal of Hydraulic Engineering* 134(7): 892–904.

Davis, W.M.. 1899. The geographical cycle. *Geographical Journal* 14: 481–504.

Diplas, P. and Vigilar, G. 1992. Hydraulic geometry of threshold channels. *Journal of Hydraulic Engineering* 118: 597–614.

Duan, J.G. and Julien, P. 2010. Numerical simulation of meandering evolution. *Journal of Hydrology* 391: 34–46.

Dulal, K.P, Kobayaski, K. and Shimizu, Y. 2010. Numerical computation of free meandering channels with the application of slump blocks on the outer bends. *Journal of Hydro-Environment Research* 3: 239–246.

Eaton, B.C. 2006. Bank stability analysis for regime models of vegetated gravel bed rivers. *Earth Surface Processes and Landforms* 31: 1438–1444.

Eaton, B.C. and Church, M. 2007. Predicting downstream hydraulic geometry: a test of rational regime theory. *Journal of Geophysical Research, Earth Surface* 112: F03025.

Edmonds, D.A. and Slingerland, R.L. 2008. Stability of delta distributary networks and their bifurcations. *Water Resources Research* 44, W09426.

Edmonds, D.A. and Slingerland, R.L. 2010. Significant effect of sediment cohesion on delta morphology. *Nature Geoscience* 3: 105–109.

Ferguson, R. and Church, M. 2009. Gravel load and aggradation in lower Fraser River. *Water Resources Research* 45: W11424.

Finnegan, N.J. and Dietrich, W.E. 2011. Episodic bedrock strath terrace formation due to meander migration and cutoff. *Geology* 39: 143–146.

Frascati, A. and Lanzoni, S. 2010. Long-term river meandering as a part of chaotic dynamics? A contribution from mathematical modeling. *Earth Surface Processes and Landform*, 35: 791–802.

French, R.H. 1985. *Open Channel Hydraulics*, New York: McGraw-Hill, 705pp.

Furbish, D.J. 1988. River-bed curvature and migration: how are they related? *Geology* 16: 752–755.

Furbish, D.J. 1991. Spatial autoregressive structure in meander evolution. *Geological Society of America Bulletin* 103: 1576–1588.

Garcia, M.H. 2007. Sediment transport and morphodynamics. In: Garcia, M., ed., *Review of Sedimentation Engineering – Processes, Measurements, Modeling, and Practice*, ASCE Manuals and Reports on Engineering Practice No. 110, Reston, VA: American Society of Civil Engineers, pp. 21–164.

Gross, L.J. and Small, M.J. 1998. River and floodplain process simulation for subsurface characterization. *Water Resources Research* 34: 2365–2376.

Guneralp, I. and Rhoads, B.L. 2011. Influence of floodplain erosional heterogeneity on planform complexity of meandering rivers. *Geophysical Research Letters* 38: L14401.

Hammer, T.R. 1973. Stream channel enlargement due to urbanization. *Water Resources Research* 8: 1530–1540.

Hancock, G.S., Small, E.E. and Wobus, C. 2011. Modeling the effects of weathering on bedrock-floored channel geometry. *Journal of Geophysical Research* 116: F03018.

Hasegawa, K. 1977. Computer simulation of the gradual migration of meandering channels. *Proceedings of the Hokkaido Branch, Japan Society of Civil Engineering* 197–202 (in Japanese).

Hey, R. and Thorne, C. R. 1986. Stable channels with mobile gravel beds. *Journal of Hydraulic Engineering* 112: 671–689.

Hoey, T.B. and Ferguson, R. 1994. Numerical simulation of downstream fining by selective transport in gravel-bed rivers: model development and illustration. *Water Resources Research* 30: 2251–2260.

Hooke, R.L. 1994. On the efficacy of humans as geomorphic agents. *GSA Today* 4: 217–225.

Howard, A.D. 1992. Modelling channel migration and floodplain sedimentation in meandering streams. In: Carling, P.A. and Petts, G.E., eds., *Lowland Floodplain Rivers, Geomorphological Perspectives*, New York: John John Wiley & Sons, pp. 1–41

Howard, A.D. 1996. Modelling channel evolution and floodplain morphology. In: Anderson, M.G., Walling, D.E. and Bates, P.D., eds., *Floodplain Processes*, Chichester: John Wiley & Sons, pp. 15–62.

Howard, A.D. and Knutson, T.R. 1984. Sufficient conditions for river meandering: a simulation approach. *Water Resources Research* 20: 1659–1667.

Huang, H.Q. and Nanson, G.C. 2000. Hydraulic geometry and maximum flow efficiency as products of the principle of least action. *Earth Surface Processes and Landforms* 15: 1–16.

Huang, H.Q., Chang, H.H. and Nanson, G.C. 2004. Minimum energy as the general form of critical flow and maximum flow efficiency and for explaining variations in river channel pattern. *Water Resources Research* 40: W04502.

Ikeda, S. and Izumi, N. 1990. Width and depth of self-formed straight gravel river with bank vegetation. *Water Resources Research* 26: 2353–2364.

Ikeda, S., Parker, G. and Sawai, K. 1981. Bend theory of river meanders. *Part I. Linear development. Journal of Fluid Mechanics* 112: 363–377.

Ikeda, S., Parker, G. and Kimura, Y. 1988. Stable width and depth of straight gravel rivers with heterogeneous bed materials. *Water Resources Research* 24: 713–722.

Johannesson, H. and Parker, G. 1989. Linear theory of river meanders. In: Ikeda, S. and Parker, G., eds., River Meandering, *Water Resources Monograph 12*, Washington, DC: American Geophysical Union, pp. 181–214.

Karssenberg, D. and Bridge, J.S. 2008. A three-dimensional numerical model of sediment transport, erosion and deposition within a network of channel belts, floodplain and hill slope: extrinsic and intrinsic controls on floodplain dynamics and alluvial architecture. *Sedimentology* 55: 1717–1745.

Kirkby, M.J. 1977. Maximum sediment efficiency as a criterion for alluvial channels. In: Gregory, K.J., ed., *River Channel Changes*, Chichester: John Wiley & Sons, pp. 429–442.

Kleinhans, M.G. 2010. Sorting out river channel patterns. *Progress in Physical Geography* 34: 287–326.

Kleinhans, M.G., Jagers, H.R.A., Mosselman, E. and Sloff, C.J. 2008. Bifurcation dynamics and avulsion duration in meandering rivers by one-dimensional and three-dimensional models. *Water Resources Research* 44: W08454.

Kleinhans, M.G., Cohen, K.M, Hoekstra, J. and IJmker, J.M. 2011. Evolution of a bifurcation in a meandering river with adjustable channel widths, Rhine delta apex, The Netherlands. *Earth Surface Processes and Landforms* 36: 2011–2027.

Knighton, D. 1998. *Fluvial Forms and Processes*, London: Hodder Arnold, 383pp.

Knox, J.C. 1976. Concept of the graded stream. In: Melhorn, W. and Flemal, R., eds., *Theories of Landform Development*, London: Allen and Unwin, pp. 168–198.

Lane, E.W. 1957. *A Study of the Shape of Channels Formed by Natural Streams Flowing in Erodible Material*, Missouri River Division Sediment Series No. 9, Omaha, NE: US Army Engineer Division, Missouri River Corps of Engineers, 106pp.

Langbein, W.B. and Leopold, L.B. 1966. *River Meanders – Theory of Minimum Variance*, US Geological Survey Professional Paper 422-H, Washington, DC: US Government Printing Office.

Larsen, E.W. and Greco, S.E. 2002. Modeling channel management impacts on river migration: a case study of Woodson Bridge State Recreation Area, Sacramento River, California, USA. *Environmental Management* 30: 209–224.

Leopold, L.B. 1973. River channel change with time. *Geological Society of America Bulletin* 84: 1845–1860.

Leopold, L.B. and Maddock, T. 1953. *The Hydraulic Geometry of Stream Channels and Some Physiographic Implications*, US Geological Survey Professional Paper 252, Washington, DC: US Government Printing Office, 57pp.

Li, S.S. and Millar, R.G. 2011. A two-dimensional morphodynamic model of gravel-bed river with floodplain vegetation. *Earth Surface Processes and Landforms* 36: 190–202.

Lisle, T.E., Pizzuto, J.E., Ikeda, H., Iseya, F. and Kodama, Y. 1997. Evolution of a sediment wave in an experimental channel. *Water Resources Research* 33: 463–512.

Luchi, R., Zolezzi, G. and Tubino, M. 2011. Bend theory of river meanders with spatial width variations. *Journal of Fluid Mechanics* 681: 311–339.

Luchi, R., Bolla Pittaluga, M. and Seminara, G. 2012. Spatial width oscillations in meandering rivers at equilibrium. *Water Resources Research* 48: W05551.

Mackin, J.H. 1948. Concept of the graded river. *Geological Society of America Bulletin* 59: 463–512.

Merritts, D.A., Walter, R., Rahnis, M., Hartranft, J., Cox, S., Gellis, A., *et al.* 2011. Anthropocene streams and base-level controls from historic dams in the unglaciated mid-Atlantic region, USA. *Philosophical Transactions of the Royal Society A* 369: 976–1009.

Murray, A.B. and Paola, C. 1994. A cellular model of braided rivers. *Nature* 371: 54–57.

Murray, A.B. and Paola, C. 2003. Modelling the effect of vegetation on channel pattern in bedload rivers. *Earth Surface Processes and Landforms* 28: 131–143.

Nanson, G.C. 1986. Episodes of vertical accretion and catastrophic stripping: a model of disequilibrium flood-plain development. *Geological Society of America Bulletin* 97: 1467–1475.

Nanson, G.C. and Croke, J.C. 1992. A genetic classification of floodplains. *Geomorphology* 4: 459–486.

Nelson, P.A. and Seminara, G. 2012. A theoretical framework for the morphodynamics of bedrock channels. *Geophysical Research Letters* 39: L06408.

Nicholas, A.P., Walling, D.E., Sweet, R.J. and Fang, X. 2006. New strategies for upscaling high-resolution flow and overbank sedimentation models to quantify floodplain sediment storage at the catchment scale. *Journal of Hydrology* 329: 577–594.

Nittrouer, J.A., Mohrig, D., Allison, M.A. and Peyret, A.-P. 2011. The lower-most Mississippi River – a mixed bedrock-alluvial channel. *Sedimentology* 58: 1914–1934.

Paola, C. 2000. Quantitative models of sedimentary basin filling. *Sedimentology* 47(Suppl. 1): 121–178.

Parker, G. 1976. On the cause and characteristic scales of meandering and braiding in rivers. *Journal of Fluid Mechanics* 76: 457–480.

Parker, G. 1978a. Self-formed rivers with stable banks and mobile bed. *Part I. The sand silt river. Journal of Fluid Mechanics* 89: 109–126.

Parker, G. 1978b. Self-formed rivers with stable banks and mobile bed. *Part II. The gravel river. Journal of Fluid Mechanics* 89: 127–148.

Parker, G. 1979. Hydraulic geometry of active gravel rivers. *Journal of Hydraulic Engineering* 105: 1185–1201.

Parker, G. 2008. Transport of gravel and sediment mixtures. In: Garcia, M., ed., *Review of Sedimentation Engineering – Processes, Measurements, Modeling, and Practice*, ASCE Manuals and Reports on Engineering Practice No. 110, Reston, VA: American Society of Civil Engineers, pp. 165–252.

Parker, G. and Cui, Y. 1998. The arrested gravel front: stable gravel–sand transitioins in rivers. 1: Simplified analytical solution. *Journal of Hydraulic Research* 36: 75–100.

Parker, G., Wilcock, P.R., Paola, C., Dietrich, W.E. and Pitlick, J. 2007. Physical basis for quasi-universal relations describing bankfull hydraulic geometry of single-thread gravel bed rivers. *Journal of Geophysical Research* 112: F04005.

Parker, G., Shimizu, Y., Wilkerson, G.V., Eke, E.C., Abad, J.D., Lauer, J.W., Paola, C., Dietrich, W.E. and Voller, V.R. 2011. A new framework for modeling the migration of meandering rivers. *Earth Surface Processes and Landforms* 36: 70–86.

Pizzuto, J.E. 1994. Channel adjustments to changing discharges, Powder River, Montana. *Geological Society of America Bulletin* 106: 1494–1501.

Pizzuto, J. 2012. Predicting the accumulation of mercury-contaminated sediment on riverbanks – an analytical approach. *Water Resources Research* 48: W07518.

Pizzuto, J.E. and the ASCE Task Committee on Hydraulics, Bank Mechanics and Modeling of River Width Adjustment. 2008. Streambank erosion and river width adjustment. In: Garcia, M., ed., *Review of Sedimentation Engineering – Processes, Measurements, Modeling, and Practice*, ASCE Manuals and Reports on Engineering Practice No. 110, Reston, VA: American Society of Civil Engineers, pp. 387–438.

Pizzuto, J., Moglen, G., Palmer, M. and Nelson, K. 2008. Two model scenarios illustrating the effects of land use and climate change on gravel riverbeds of suburban Maryland, U.S.A. In: Habersack, H., Piégay, H. and Rinaldi, M., eds., *Gravel-bed Rivers VI: from Process Understanding to River Restoration*, Amsterdam: Elsevier, pp. 357–386.

Robinson, R.A.J. and Slingerland, R.L. 1998. Origin of fluvial grain-size trends in a foreland basin: the Pocono Foramtion of the central Appalachian Basin. *Journal of Sedimentary Research* 68: 473–486.

Roe, G.H., Montgomery, D.R. and Hallet, B. 2002. Effects of orographic precipitation variations on the concavity of steady-state river profiles. *Geology* 30: 143–146.

Sambrook Smith, G.H. and Ferguson, R. 1995. The gravel–sand transition along river channels. *Journal of Sedimentary Research* A65: 423–430.

Seminara, G. 2006. Meanders. *Journal of Fluid Mechanics* 554: 271–297.

Seminara, G. 2010. Fluvial sedimentary patterns. *Annual Reviews of Fluid Mechanics* 42: 43–66.

Sinha, S.K. and Parker, G. 1996. Causes of concavity in longitudinal profiles of rivers. *Water Resources Research* 32: 1417–1428.

Skalak, K. and Pizzuto, J. 2010. The distribution and residence time of suspended sediment stored within the channel margins of a gravel-bed bedrock river. *Earth Surface Processes and Landforms* 35: 435–446.

Sklar, L.S. and Dietrich, W.E. 2004. A mechanistic model for river incision into bedrock by saltating bed load. *Water Resources Research* 40: W06301.

Slingerland, R., Harbaugh, J.W. and Furlong, K.P. 1994. *Simulating Clastic Sedimentary Basins*, Englewood Cliffs, NJ: Prentice-Hall, 220pp.

Snow, R.S. and Slingerland, R.L. 1987. Mathematical modeling of graded river profiles. *Journal of Geology* 95: 15–33.

Stock, J.D. and Dietrich, W.E. 2006. Erosion of steepland valleys by debris flows. *Geological Society of America Bulletin* 118: 1125–1148.

Stock, J.D. and Montgomery, D.R. 1999. Geologic constraints on bedrock river incision using the stream power law. *Journal of Geophysical Research* 104: 4983–4993.

Stolum, H. 1996. River meandering as a self-organization process. *Science* 271: 1710–1713.

Sun, T., Meakin, P., Jossang, T. and Schwarz, K. 1996. A simulation model for meandering rivers. *Water Resources Research* 32: 2937–2954.

Sutherland, D.G., Ball, M.H., Hilton, S.J. and Lisle, T.E. 2002. Evolution of a landslide-induced sediment wave in the Navaro River, California. *Geological Society of America Bulletin* 114: 1036–1048.

Thomas, W.A. and Chang, H. 2008, Computational modeling of sedimentation processes. In: Garcia, M., ed., *Review of Sedimentation Engineering – Processes, Measurements, Modeling, and Practice*, ASCE Manuals and Reports on Engineering Practice No. 110, Reston, VA: American Society of Civil Engineers, pp. 649–682.

Thonon, I., de Jong, K., van der Perk, M. and Middelkoop, H. 2007. Modelling floodplain sedimentation using particle tracking. *Hydrological Processes* 21: 1402–1412.

Turowski, J.M., Lague, D. and Hovius, N. 2007. Cover effect in bedrock abrasion: a new derivation and its implciations for the modeling of bedrock channel morphology. *Journal of Geophysical Research* 112: F04006.

Turowski, J.M., Hovius, N., Wilson, A. and Horng, M-J. 2008. Hydraulic geometry, river sediment and the definition of bedrock channels. *Geomorphology* 99: 26–38.

Van De Wiel, M.J., Coulthard, T.J., Macklin, M.G. and Lewin, J. 2007. Embedding reach-scale fluvial dynamics within the CAESAR cellular automaton landscape evolution model. *Geomorphology* 90: 283–301.

Vigilar, G.G. and Diplas, P. 1997. Stable channels with mobile bed: formulation and numerical solution. *Journal of Hydraulic Engineering* 123: 189–199.

Whipple, K.X. 2004. Bedrock rivers and the geomorphology of active orogens. *Annual Reviews of Earth and Planetary Sciences* 32: 151–185.

Whipple, K.X., Hancock, G. and Anderson, R. 2000. River incision into bedrock: mechanics and relative efficacy of plucking, abrasion and cavitation. *Geological Society of America Bulletin* 112: 490–503.

White, W.R., Bettess, R. and Paris, E. 1982. Analytical approach to river regime. *Journal of Hydraulic Engineering* 108: 1179–1193.

Wolman, M.G. 1955. *The Natural Channel of the Brandywine Creek*, US Geological Survey Professional Paper 271, Washington, DC: US Government Printing Office, 56pp.

Xu, D., Bai, Y., Ma, J. and Tan, Y. 2011. Numerical investigation of long-term planform dynamics and stability of river meandering on fluvial floodplains. *Geomorphology* 132: 195–207.

CHAPTER 20

Experimental studies and practical challenges in fluvial geomorphology

François Métivier[1], Chris Paola[2], Jessica L. Kozarek[2] and Michal Tal[3]
[1]Institut de Physique du Globe de Paris, Université Paris – Sorbonne Paris Cité, Université Paris Diderot, CNRS, Paris, France
[2]University of Minnesota, Minneapolis, MN, USA
[3]Université Aix-Marseille, CEREGE UMR 7330, Europôle de l'Arbois, Aix-en-Provence, France

20.1 Introduction

For much of the 20th century, river engineering focused primarily on creating hard structures aimed at control and human, often commercial, functionality. In this endeavour, experimentation with scale models played a central role, especially before the advent of inexpensive computers and techniques for using them to model flow in rivers. Since then, and especially within the arena of stream restoration, experimental work has declined in importance for practice, even as it has grown in importance for basic research. In some parts of the world (e.g. the United States), local restoration projects are often done by small operators that cannot individually afford experimental studies; as an enterprise, river management and restoration have been relatively unorganized and thus lack a means of motivating and focusing experimental research. Furthermore, management often involves biotic and/or geochemical aspects that are not part of traditional riverine experimental programmes.

Nevertheless, there is a strong case to be made for using experiments as part of an overall effort to improve the scientific basis for river management. Experimental research on stream dynamics generally has blossomed in recent years and these improvements in understanding and prediction of the natural behaviour of streams contribute to the scientific underpinnings of river practice and management. However, with the exception of a recent review by Rice *et al.* (2010) that focused on experiments with live organisms, the potential of experimental studies focused on stream management and restoration has barely been tapped. In this chapter, we provide some examples of experimental studies especially relevant to management and restoration, focusing on geomorphology and physical processes. We summarize our views on scaling and other issues related to using experiments to their best advantage and suggest themes for future experimental research.

We begin with a comment about strategy and motivation. Broadly, the two strategies that one might adopt for experimentation could be thought of as (i) a 'model study' approach in which one builds a scaled model of a particular restoration site and tests various strategies in it and (ii) experimental programmes aimed not at specific sites but at generic questions and problems. Either approach is feasible, but our emphasis here will be on the latter, if for no other reason than that most small-scale practical projects cannot afford the costs of adding a site-specific model study to their design process. It is possible, however, that stricter standards for specific outcomes in management and restoration (i.e. clearly defined objectives with penalties for failing to meet them) would provide the motivation for additional experimental testing of proposed designs.

Focusing on the second strategy, what are the benefits of experimental studies for stream management and restoration? The obvious one is evaluation of proposed methods under controlled conditions. There is no question that the ultimate test of any technique or structure is how it performs in the field, but this is influenced by many factors, some extrinsic to the technique itself. Even for a technique that has generally worked well in the field, there is value in studying it under fully controlled conditions, understanding better how it works and how it fails and, from that, potentially finding ways to improve the design or predict its limitations better. Another benefit of experimental studies is the possibility of studying long-term behaviour that can only be predicted or inferred in the field. This is important because many processes in natural systems evolve over time-scales well beyond the typical time frame of a grant or monitoring programme and the insight from long-term observation is missed. Finally, because stream-manipulation techniques often simultaneously influence several variables affecting channel dynamics, the control afforded by laboratory experiments can be instrumental in evaluating the role of individual variables.

Experiments in geology and geomorphology date back to the late 19th to early 20th century. Experimentalists such as Daubrée (1879) and Jaggar (1908) had the intuition that physical models were somehow mimicking the real world in a way that might raise new insight and questions, if not answers. During the succeeding years, experiments related to geomorphological questions progressively evolved along two more or less distinct

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

paths. One of these paths was that of hydraulic engineers who focused on scale problems and ways to address these issues in small-scale models. The reference book by Yalin (1971) is the quintessence of such an approach. Others, in the footsteps of Schumm (1977) and Schumm *et al.* (1987), worked on geomorphological analogue models, trying to reproduce diverse features such as meanders, braids, alluvial fans and terraces. These models were meant to show the evolution through time of simplified systems and inspire hypotheses to unravel the dynamics behind the landscapes. From the 1980s on, these approaches progressively converged to produce 'Froude scaled' models of alluvial plains (Ashmore 1985, 1991; Ashworth *et al.* 1994). The turn of the 21st century corresponds to the development of microscale (order of 1 m) models (Coleman and Eling 2000; Métivier and Meunier 2003; Lague *et al.* 2003). This development coincides with physical studies showing that hydraulically scaled models are not necessary to achieve predictability and that models using laminar flows are capable of providing quantitative insights into natural processes operating in the fluvial system (Malverti *et al.* 2008; Paola *et al.* 2009; Lajeunesse *et al.* 2010b). Today, researchers and stakeholders can choose among a range of facilities and techniques ranging from the smallest flume to the largest outdoor facility and study a wide variety of problems in fluvial morphology. Here we discuss their respective interest.

20.2 Experimental methods and facilities

Basic equipment

Stream experimentation typically centres on either a flume (Fig. 20.1a), a relatively long, narrow experimental channel with fixed parallel side walls, or a stream table (Fig. 20.1b), a relatively wide, shallow basin in which the experimental river can find its own width and channel pattern. These facilities are in themselves extremely simple and require no sophisticated technology. Either type may be mounted on a tilting structure so that the slope may be adjusted. If not, the sediment surface will find its own transport slope for given imposed water and sediment discharges, although for low transport rates this can take some time. Refinements come in when we turn to measurement. Typically one wants to at least measure topography and the flow field, the latter in two or three dimensions and including if possible measurement of turbulence.

(a)

(b)

Figure 20.1 (a) Typical tilting flume. Source: Craig Hill, Senior Research Associate, St Anthony Falls Laboratory. and (b) typical stream table, both from St Anthony Falls Laboratory. The flume is being used to study scour around rock structures and the stream table to study braiding.

Topography

Currently, there are two general approaches to measuring topography: scan-based and image-based (local measurement by hand with a point gauge is rarely performed any longer). An example of a scanner developed at St Anthony Falls Laboratory at the University of Minnesota (SAFL) is shown in Fig. 20.2. The scan is done from a cart equipped with a precise, fast positioning system and a scanning device that sweeps over the surface, recording elevation as it goes. There are tradeoffs involving speed, dynamic range, laser cost, resolution, and so on, but a typical example setup measures a 600 mm swath at 250 mm s^{-1} on a 1 × 1 mm grid (150,000 points s^{-1}) with 0.3 mm vertical resolution.

An equally effective alternative to point scanning is image-based measurement (for a review of imagery-based techniques, see Tal *et al.* 2012). This can be done using conventional photogrammetry, although it is often difficult to obtain the necessary resolution at small scales. Recently, a particularly

Figure 20.2 An automated topography scanner from St Anthony Falls Laboratory capable of measuring 150,000 points s^{-1} to a resolution of 0.3 mm. The scanner is shown imaging the scour pattern around a rock structure similar to those installed in numerous restoration projects, from the channel shown in Fig. 20.1. Source: Craig Hill, Senior Research Associate, St Anthony Falls Laboratory.

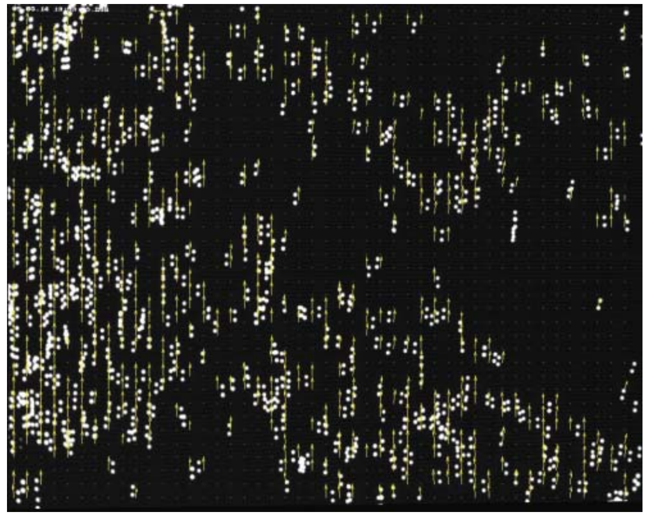


Figure 20.3 Particle-image velocimetry (PIV) provides high-resolution flow fields based on tracking movement of particles between successive images. This realization of a PIV surface velocity vector field includes photographs of both of the pair of corrected raw images of floating confetti used to do the PIV analysis as well as the resulting velocity vectors. Source: Chris Ellis, Senior Research Associate, St Anthony Falls Laboratory.

sophisticated yet affordable method has been developed using moiré techniques that has proven very useful in geomorphology laboratory experiments (Limare *et al.* 2011; Tal *et al.* 2012). Examples include the study of submarine channel formation and microscale braided rivers. In all these examples a simple method consisting of grid projection combined with Fourier transform analysis was used. This method can be greatly elaborated through the use of a phase shifting method for phase calculation, robust phase unwrapping combined with grey coding and a calibration procedure that allows the assessment of the geometric parameters. This system is relatively easy to use and calibrate and can result in very high-resolution acquisition of both topography and flow depth for even millimetre-scale channels (Limare *et al.* 2011; Tal *et al.* 2012).

Flow field

Traditionally, water flow has been measured at a point using one of a variety of methods (e.g. micro-propeller, Pitot tube). Acoustic Doppler velocimeters (ADVs) are now relatively inexpensive and easy to use, although their size typically makes them impossible to use in channels shallower than a few centimetres. An alternative to any of these point- or profile-based methods is particle image velocimetry (PIV), which amounts to determining velocity from successive images of particles moving with the fluid (Tal *et al.* 2012). In flows at least a few centimetres deep this is traditionally done using vertical laser light sheets to resolve turbulence structure. In shallow experimental flows, PIV can instead be performed in plan view by imaging particles moving on the surface of the flow (Fig. 20.3).

Facilities

Experimental facilities for the study of stream dynamics have been developing rapidly around the world. Whereas historically such facilities were housed mainly in engineering laboratories and departments, today they are increasingly becoming a part of Earth-science ones as well. Table 20.1 summarizes some of the existing experimental facilities focused on stream dynamics. Below we present some specific examples of ongoing programmes and capabilities at the facilities where we work.

SAFL and Outdoor StreamLab

St Anthony Falls Laboratory, University of Minnesota, Minneapolis, is one of the larger experimental facilities in the world for all aspects of fluid flow and its relation to sediment, Earth science, biology and medicine. Throughout its history, SAFL has hosted visitors from around the world to take advantage of its people and facilities. The growth of the Internet now makes it possible to run experiments remotely, with minimal intervention by local staff. For example, the measurement system shown in Fig. 20.2 can be programmed remotely to collect a variety of data – flow velocity, topography, etc. – over any spatial grid and time interval desired. An example of the detailed measurements of flow and turbulence that such experiments can provide is outlined in the section 'Instream structures' below.

Stream management frequently includes ecological objectives, which can be difficult to study at reduced scales (Orr *et al.* 2009; Wilcock *et al.* 2008). One approach to this problem is to set up instrumented watersheds in which manipulations (e.g. addition of in-channel wood) can be carried out and their effects evaluated.

Another approach to maintaining the scientific value of controlled experimentation while allowing for the development of natural ecosystems is the construction of outdoor controlled experimental facilities in which ecosystem processes

Table 20.1 List of facilities where experiments in fluvial geomorphology can be performed. This list is a simplified version of a more complete list maintained at https://morpho.lpgp.fr/OSS/Facilities.

Country	Laboratory	Contact	Experiment types	Reference
Canada	Mountain Channel Hydraulic Experimental Laboratory (MCHEL), University of British Columbia	Marwan Hassan	Multiple instrumented flumes for research in steeplands rivers and sediment transport	Zimmermann, A.M., Church, M. and Hassan, M.A. 2010. Step-pool stability: testing the jam state hypothesis. *Journal of Geophysical Research* 115: F02008
Canada	Newalta Environmental Fluids Laboratory – River Modeling Flume, at University of Western Ontario	Peter Ashmore	Physical model, gravel bed, channel pattern, river morphology, photogrammetry	Egozi, R. and Ashmore, P. 2009. Experimental analysis of braided channel pattern response to increased discharge. *Journal of Geophysical Research* 114: F02012
Canada	Civil Engineering Hydraulics Laboratory (CEHL), University of Ottawa	Colin D. Rennie	Bend flow, bank erosion, turbulence, acoustic velocimetry	Jamieson, E., Post, G. and Rennie, C.D. 2010. Spatial variability of three dimensional Reynolds stresses in a developing channel bend. *Earth Surface Processes and Landforms* 35: 1029–1043
France	Laboratory for Hydraulics Saint Venant (LHSV)	Damien Pham Van Bang	Meanders, bank erosion	
France	Géosciences Rennes	Dimitri Lague	Catchment and landscape development, fluvial incision, orographic precipitation	Lague, D., Crave, A. and Davy, P. 2003. Laboratory experiments simulating the geomorphic response to tectonic uplift. *Journal of Geophysical Research, Solid Earth* 108(B1): ETG3-1–ETG3-20
France	Geological Fluid Dynamics, Institut de Physique du Globe de Paris	E. Lajeunesse, F. Métivier	Microscale models, fluvial morphology, sediment transport	Métivier, F., Paola, C., Kozarek, J. and Tal, M. Experimental studies and practical challenges in fluvial geomorphology. In: Kondolf, M. and Piégay, H., eds., *Tools in Fluvial Geomorphology*, John Wiley & Sons, in press
France	Geosciences Montpellier, University of Montpellier, France	S. Dominguez and J. Malavieille	Orogenic wedge, accretionary prism, geomorphology, tectonic–erosion–sedimentation interactions, earthquake dynamics, fault kinematics, thrust and normal faulting	Graveleau, F., Hurtrez, J.-E., Dominguez, S. and Malavieille, J. 2011. A new experimental material for modelling relief dynamics and interactions between tectonics and surface processes. *Tectonophysics* 513(1–4): 68–87
France	Erosion Torrentielle Neige et Avalanches, IRSTEA	Alain Recking	Morphodynamics, steep slope, DEM	Recking, A., Frey, P., Paquier, A., Belleudy, P. and Champagne, J.Y. 2008. Bedload transport flume experiments on steep slopes. *Journal of Hydraulic Engineering* 134: 1302–1310
Italy	Hydraulic Laboratory, Department of Civil and Environmental Engineering, University of Trento	Walter Bertoldi	Large flume, sediment transport, braided rivers	Bertoldi, W., Zanoni, L. and Tubino, M. 2009. Planform dynamics of braided streams. *Earth Surface Processes and Landforms* 34(4): 547–557
Switzerland	Laboratory of Hydraulic Constructions, Ecole Polytechnique Fédérale, Lausanne	Koen Blanckaert	Confluence, laboratory experiments, fluid dynamics, morphology	Leite Ribeiro, M., Blanckaert, K., Roy, A.G. and Schleiss A.J. 2012. Flow and sediment dynamics in channel confluences. *Journal of Geophysical Research* 117: F01035
Switzerland	Environmental Hydraulics Laboratory, Ecole Polytechnique Fédérale, Lausanne	Koen Blanckaert	Open-channel bend, meander, laboratory experiments, fluid dynamics, morphology	Blanckaert, K. 2010. Topographic steering, flow recirculation, velocity redistribution and bed topography in sharp meander bends. *Water Resources Research* 46: W09506
UK	Total Environment Simulator	Stuart McLelland, Daniel Parsons	Total environment simulator	

(*continued overleaf*)

Table 20.1 (*continued*)

Country	Laboratory	Contact	Experiment types	Reference
USA	St Anthony Falls Laboratory	Jeff Marr	Morphodynamics, sediment transport , fluid dynamics, riparian ecology in rivers and deltas, including effects of climate and subsidence	Paola, C., Twilley, R.R., Edmonds, D.A., Kim, W., Mohrig, D., Parker, G., Viparelli, E. and Voller, V.R. 2011. Natural processes in delta restoration: application to the Mississippi Delta. *Annual Review of Marine Science* 3(1): 67–91
USA	Hydraulics Laboratory, Colorado State University	Pierre Julien	River engineering and sedimentation	Julien, P., Richard, G. and Albert, J. 2005. Stream restoration and environmental river mechanics. *International Journal of River Basin Management* 3(3): 191–202
USA	Ven Te Chow Laboratory, University of Illinois	Marcelo Garcia	Morphodynamics, sediment transport , fluid dynamics and riparian ecology in rivers, deltas and under water	Garcia, M.H., ed. 2008. *Sedimentation Engineering : Processes, Measurements, Modeling and Practice*, Reston, VA: American Society of Civil Engineers, 1132 pp.
USA	IIHR-Hydroscience and Engineering	Thanos Papanicolau	General stream dynamics, flow and sedimentation, watershed sediment dynamics, fish passage, phytoremediation	Papanicolaou, A.N., Hobbs, B., Kramer, C., Fox, J.F. and Kjos, L. 2005. Fluid–sediment dynamics around a barb: an experimental case study of a hydraulic structure for the Pacific Northwest. *Canadian Journal of Civil Engineering* 32(5): 853–867
USA	Baker Environmental Hydraulics Laboratory, Virginia Tech	Panos Diplas	Environmental, fluvial, ecological and infrastructure hydraulics	Kozarek, J., Hession, C. and Diplas, P. 2010. Hydraulic complexity metrics for evaluating in-stream brook trout habitat. *Journal of Hydraulic Engineering* 136(12): 1067–1076
USA	Center for Ecohydraulics Research, University of Idaho	Ralph Budwig	Headwater streams, mountain rivers	Tonina, D. and Buffington, J.M. 2009. Hyporheic exchange in mountain rivers. I. Mechanics and measuring hyporheic exchange. *Geography Compass* 3(3): 1063–1086.
USA	University of Texas Morphodynamics Laboratory	Joel P. Johnson	Delta evolution in response to sediment supply, tectonics and sea-level change; vegetation-delta morphodynamics; fluvial autogenic processes and allogenic responses; submarine processes and deposits; flash floods and tsunamis	Sanguinito, S. and Johnson, J. 2012. Quantifying gravel overlap and dislodgement forces on natural river bars: implications for particle entrainment. *Earth Surface Processes and Landforms* 37(1): 134–141
USA	Baylor Experimental Aquatic Research	Ryan S. King	Controlled experiments on the effects of physical, chemical and biological variables on stream ecosystems, with particular emphasis on identifying targets for management and restoration of wadeable streams	Taylor, J.M., Back, J.A. and King, R.S. 2012. Grazing minnows increase benthic autotrophy and enhance the response of periphyton elemental composition to experimental phosphorus additions. *Freshwater Science* 31: 451–462
USA	Penn Sediment Dynamics Laboratory	Doug Jerolmack	Alluvial fans, bedload transport, delta hydrodynamics	Martin, R.L., Jerolmack, D.J. and Schumer, R. 2012. The physical basis for anomalous diffusion in bedload transport. *Journal of Geophysical Research* 117: F01018

can develop but where the facility is directly associated with a fully equipped laboratory. An example of this approach is the SAFL Outdoor StreamLab (OSL) (Fig. 20.4).

The OSL allows for near full-scale stream experiments under ambient sunlight and weather conditions that include major ecosystem components from microorganisms to macroinvertebrates and fish with laboratory-quality measurement and control. It uses an abandoned spillway adjacent to St Anthony Falls on the Mississippi River and to SAFL. The currently active part of the OSL, called the Riparian Basin, is an approximately 20 × 40 m area where the limestone spillway was filled with ~ 1.5–2 m of sandy material bounded by a concrete headbox and tailbox and surrounded by plastic interlocking sheet piles to minimize groundwater leakage and allow for the study of surface water–groundwater interactions. Water flows into the facility from the Mississippi River under valve control and is constantly monitored using an air ultrasonic transducer to measure water level over a contracted weir (Fig. 20.4). Sediment is fed into the channel using a variable-speed auger just downstream of the weir and sediment rates are verified by periodic grab samples. Water and sediment flow through the system to the settling basin in the tailbox, where sediment is collected, measured by total station survey and recirculated for re-use. The OSL has been designed for flexibility: for example, different channel and floodplain configurations can be examined.

The OSL at present was configured as a small sand-bed meandering stream channel (bankfull width ~ 2.7 m, depth ~0.2 m and sinuosity ~1.3) in June 2008 using coconut-fibre matting to fix the bank geometry. This channel size allows for flood-plain inundation in the range 1–2 cm for the maximum

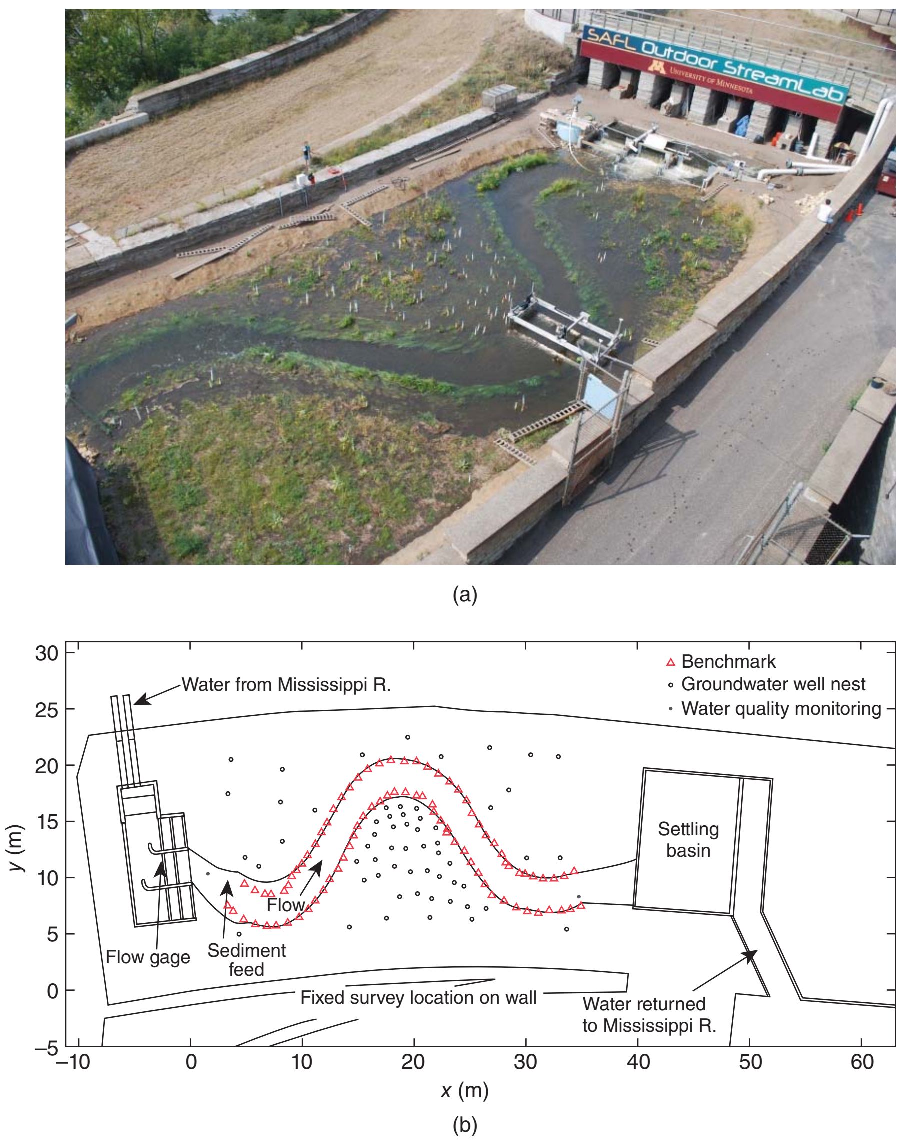


Figure 20.4 (a) Photograph and (b) sketch layout of the SAFL Outdoor StreamLab (OSL) with measurement/monitoring locations. Source: Chris Ellis, Senior Research Associate, St Anthony Falls Laboratory.

discharge available. The banks and floodplain were planted with Minnesota-native herbaceous vegetation. Between meanders, riffles were constructed from a mixture of cobble and course-grained material to mimic natural pool–riffle geometry and to provide a stable substrate for colonization by instream biota. With the exception of the riffle areas, the channel bed is a mobile coarse-grained sand mixture (median grain size, $D_{50} = 0.7$ mm). The channel was constructed with a flat bed, but within the first flood event bars formed near the inner bank of the second and third meander bends and persisted through the subsequent four seasons of experiments. The average bed slope following these first floods in July 2008 was 0.007. Experimental water flow rates in this channel ranged from very low flow (~ 0.025 m^3 s^{-1}) to bankfull flow (~ 0.280 m^3 s^{-1}) to large overbank floods (~ 1.2 m^3 s^{-1}). Sediment feed rates ranged from no sediment feed to high (~ 7 kg min^{-1}) to very high values (~ 20 kg min^{-1}), representing flood events with high water and sediment feed separated by periods with low flow and sediment feed. This strategy allows for controlled full-scale experiments on the physical, chemical and biological interactions among a channel, its floodplain and vegetation (Rominger *et al.* 2010).

In addition to flow and sediment measurement and control systems, a major advantage of the OSL experimental facilities is high-resolution measurement capabilities. A portable data acquisition (DAQ) cart similar to indoor instrumentation carriages designed and built at SAFL (Fig. 20.5) is available to collect spatially distributed data. The DAQ cart can scan an area of approximately 1.3 $\times$ 3 m to sub-centimetre resolution. To scan the entire stream channel, multiple cart stations are stitched together with sub-millimetre accuracy using conical benchmarks distributed along the stream channel. This process, although labour intensive, results in centimetre resolution channel and water-surface topography; examples are given in the section 'Instream structures' below. One-dimensional traverses, located by survey measurements, are used to take additional high-resolution, spatially referenced velocity measurements using an ADV. Groundwater flow is measured using a network of piezometer nests installed in 2008 (Nowinski *et al.* 2011). Continuous monitoring of water quality (temperature, dissolved oxygen, turbidity, conductivity and pH) upstream and downstream, an on-site weather station and a time-lapse camera provide supplementary data for OSL experiments. Future OSL plans include larger DAQ carts and the development of a second, longer (120 m) stream channel.

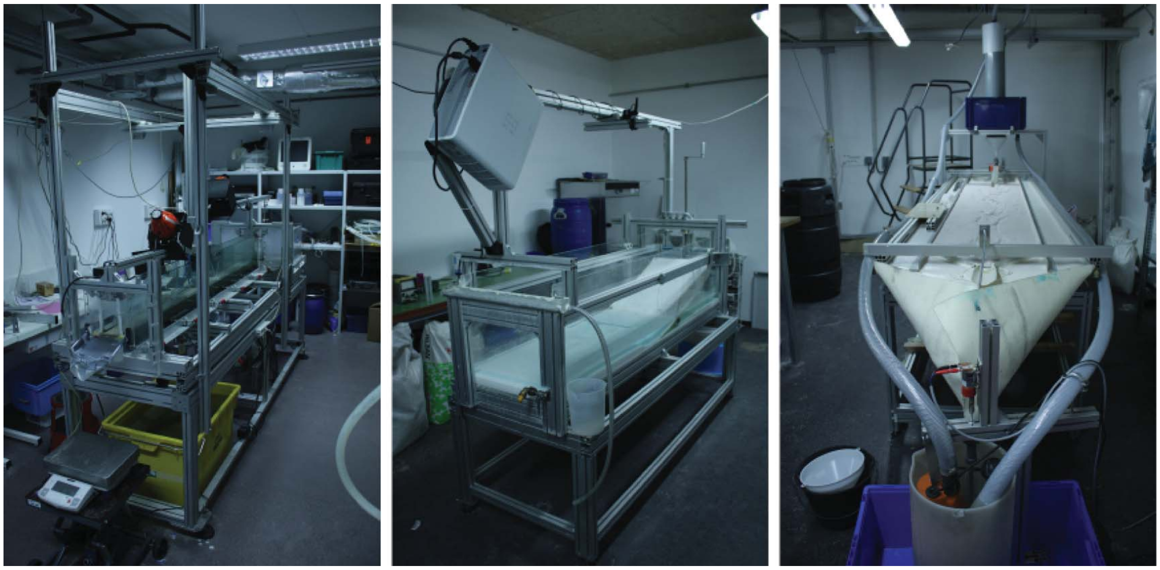
Figure 20.5 Portable data acquisition (DAQ) cart (1.3 × 3 m) outfitted with an ultrasonic transducer to measure water surface, sonar to measure subaqueous topography and a downward-looking laser to measure above water topography. Individual cart locations are located using benchmarks and stitched together. Photograph courtesy Anne Lightbody, University of New Hampshire.

This combination of experimental control over channel topography, sediment feed and water flow rate, in a nearly full-scale experimental channel with natural weather and sunlight, allows the study of ecological processes that are difficult or impossible to scale (Wilcock *et al.* 2008). For example, a study by Merten *et al.* (2010) examined the influence of velocity and bedload transport on periphyton accrual in the presence of macroinvertebrate grazers, processes that depend on sunlight and river water chemistry in addition to physical stream characteristics. A study by Rominger *et al.* (2010) investigated the changes to flow structure and topography induced by vegetation planted on a point bar in the meandering channel in the OSL, and Nowinski *et al.* (2011) quantified the effect of temporally changing hydraulic conductivity (which has implications for ecological and biogeochemical processes) within the OSL floodplain. The OSL is uniquely suited to study interactions between physical, chemical and biological processes in both surface and subsurface flows at a range of spatial scales from microorganism to reach scale.

A major part of the value of the OSL is that, as part of a large laboratory, it is closely integrated with traditional experiments in addition to numerical models. The SAFL philosophy is to view the OSL as one vertex in a triad of StreamLabs – Outdoor, Indoor and Virtual – that together provide a comprehensive approach for tackling stream-restoration research.

Microscale experiments: IPGP

The University of Paris Diderot facility at IPGP is a good example of a laboratory that includes a range of experimental stream facilities in a small space. Three types of experiments are performed at the IPGP laboratory. Small-scale flumes (Fig. 20.6a) are used to study the dynamics of particle transport or bedform migration and formation. A simple 'bathtub' or tank (1 $\times$ 0.5 $\times$ 0.5 m; Fig. 20.6b) fitted with an inclined floor allows the study of subaquatic processes such as the dynamics of channel inception by density currents. Stream tables of varying sizes (2 $\times$ 0.7 m on average; Fig. 20.6c) allow the reproduction of braided streams, alluvial fans or self-formed single-thread channels. The scale of these experiments, on the order of several metres, makes it possible to develop an entire laboratory within a small area (less than 200 m^2). The small size affords additional advantages: reduced experiment time, a high level of control and reduced cost of materials. There is a strong relation between system size and intrinsic time-scales (e.g. time to reach steady



(a) (b) (c)

Figure 20.6 IPGP facilities for small-scale experiments: (a) flume, (b) tank and (c) stream table.

state), so a reduction in size means that an experiment lasts from a few hours up to at most a few days. Most of the experiments are set up and run within a working day. The discharge is easily controlled and does not require large pumps or tanks. Because the volume of sediment in these experiments is small, it is easy to perform tests and carry out studies using a wide range of material.

Microscale experiments are thus emerging as a valuable complement to larger experiments and other approaches, as they offer the possibility, at very reasonable cost, of exploring ranges in parameters in a very short period, allowing researchers to tackle problems that could take years to study otherwise. Over the past decade, small-scale experiments have attracted increasing attention and are being used and developed in a growing number of laboratories worldwide. Extensive discussion of the use and possibilities offered by such experiments can be found in Lajeunesse *et al.* (2010a). Scaling issues will be discussed later in this chapter, but in most cases they can be circumvented if the goals are well thought out and appropriate for reduced-scale experimentation.

Other facilities

SAFL and IPGP are two examples of facilities where experimental fluvial geomorphology can be performed for both basic and applied research using a variety of model types. There are numerous laboratories where practitioners can find experienced researchers and facilities to develop collaborations. Table 20.1 summarizes some of these. An updated list with supplementary information can be found at https://morpho.ipgp.fr/OSS/Facilities. Information relating to laboratory experiments in sedimentology and stratigraphy can also be found at http://sedimentexperiments.blogspot.fr/.

20.3 Example experimental studies

Wood in rivers

Studies of wood in rivers have had a resurgence in recent years spurred by management and restoration seeking to balance the often conflicting needs of society (navigation, recreation) and public safety (flood risk, damage to infrastructure) – which mandate removal of woody debris – and ecology (channel complexity, aquatic habitat) – which depends on wood recruitment and retention (Moulin and Piégay 2004). The influence of wood on channel dynamics is varied and complex. Wood introduced to a river obstructs the flow and alters the channel hydraulics, exerting a first-order control on channel morphology across a wide range of scales from channel roughness and surface grain size to the creation of in-channel features and channel patterns and the formation of floodplains and valley bottom landforms (Keller and Swanson 1979; Nakamura and Swanson 1993; Lisle 1995; Abbe and Montgomery 1996; Manga and Kirchner 2000; Gurnell *et al.* 2002; Montgomery *et al.* 2003). Wood is often associated with reduction and deflection of flow paths and velocity (Huang and Nanson 1997; Bennett *et al.* 2002, 2008) and riparian forests are more often associated with bank stability enhancement and channel narrowing (Andrews 1984; Hey and Thorne 1986). However, if the size of the stream is small, the influence of vegetation can lead to stream widening (Zimmerman *et al.* 1967; Hession *et al.* 2003). Because of the various consequences of in-channel wood, it is important for river managers to consider thoroughly how to best manage it.

Much of our understanding of the behaviour of wood in rivers is based on studies in relatively pristine systems with old-growth forests where the trees are very large compared with the size of individual channels; or localized geomorphic investigations around debris jams that can remain in the channel for up to several decades (Abbe and Montgomery 1996). The most important impact of historical river and riparian forest management has been a reduction in the quantity and size of wood input, leading to decreased wood storage and increased mobility (Gurnell 2003). Therefore, many river management and restoration questions pertain to the transport dynamics of wood through a reach and in particular the time-scales characterizing its transfer and storage.

The complex and ambiguous interaction between channel morphodynamics and in-channel wood coupled with the challenges of tracking wood across large spatial scales and over long time-scales in the field make it a good target for experimental study (e.g. Brauderick and Grant 2000, 2001; Bocchiola *et al.* 2006; Wilcox and Wohl 2006; Jiang *et al.* 2009).

A microscale experimental study at the IPGP facility was designed to study braiding morphodynamics and wood transport and storage. A braided morphology was chosen because braided rivers flow across wide alluvial plains which have a large storage potential for wood. In addition, braid plains are reworked by progressive lateral channel migration through bank erosion in addition to spontaneous channel switching, which activates new channels and abandons old ones, thus allowing the study of a broad range of interactions between wood and channel dynamics. Finally, braided patterns are extremely simple to reproduce in the laboratory (more on this is given in the section 'Vegetation, fine sediment and the quest for laboratory-scale meandering' below). The experiments were conducted in a stream table that was 2 m long and 0.75 m wide with an adjustable slope set to 0.05. The flow was laminar (Q_w = 1.5 L min^{-1}) and sediment was input at a constant rate (Q_s = 12 g min^{-1}). Once a fully braided morphology had developed to a dynamic equilibrium (sediment input equalled sediment output), the flow was shut off and small plastic particles (3 × 2 mm) used to simulate logs were dispersed randomly over the entire braid plain (Fig. 20.7). The ratio of the density of the logs to the density of water was ∼1, so the transport of logs was highly sensitive to flow conditions. The experiments were set up to simultaneously measure log movement using continuous time-lapse imagery (30 s intervals) and bed topography and flow depth acquired every 10 min using the moiré technique discussed above. Time-lapse images were processed to determine the number of logs remaining on the braid plain through time and also track the exact positions of individual logs. Work is in progress to couple log dynamics to a full range of statistics describing bed topography and flow.

Preliminary results demonstrate a strong link between the removal time of logs in the braid plain and the intrinsic reworking time of the braid plain (by channel migration, avulsion and bifurcation), which is proportional to the average cross-sectional area of the flow and inversely proportional to the sediment supply (Cazanacli *et al.* 2002). Figure 20.8(a) shows the cumulative fraction of an experimental braid plain that was visited (i.e. reworked) by the flow through time for steady-state braiding. At the beginning of an experiment (t = 0), ∼40% of the braid plain was inundated and 60% of the bed was dry. Channels reworked 85% of the braid plain after ∼ 4 h. A long time was required for the remaining 15% of the braid plain to be visited by the flow. The consequence of this intrinsic behaviour for wood present in the braid plain is that logs in or near the active flow paths are removed from the system quickly, whereas logs in those parts of the braid plain that remain unvisited by flow for long periods remain in place. The number of logs present on the braid plain through time for a run shows two distinct phases (Fig. 20.8b). In the first phase, logs that are immediately in the flow paths are quickly flushed out of the system. This phase strongly depends on the initial distribution of logs that were randomly seeded. In the second phase, the number of logs decays logarithmically with time (i.e. the rate at which logs are removed from the braid plain is inversely proportional to time), reflecting the long time

Figure 20.7 Negative image of microscale experimental braidplain with plastic particles (in white) used to simulate logs dispersed randomly across the bed at the start of an experiment. Flow is from top to bottom. Scale along left edge denotes 1 cm and 5 cm increments.

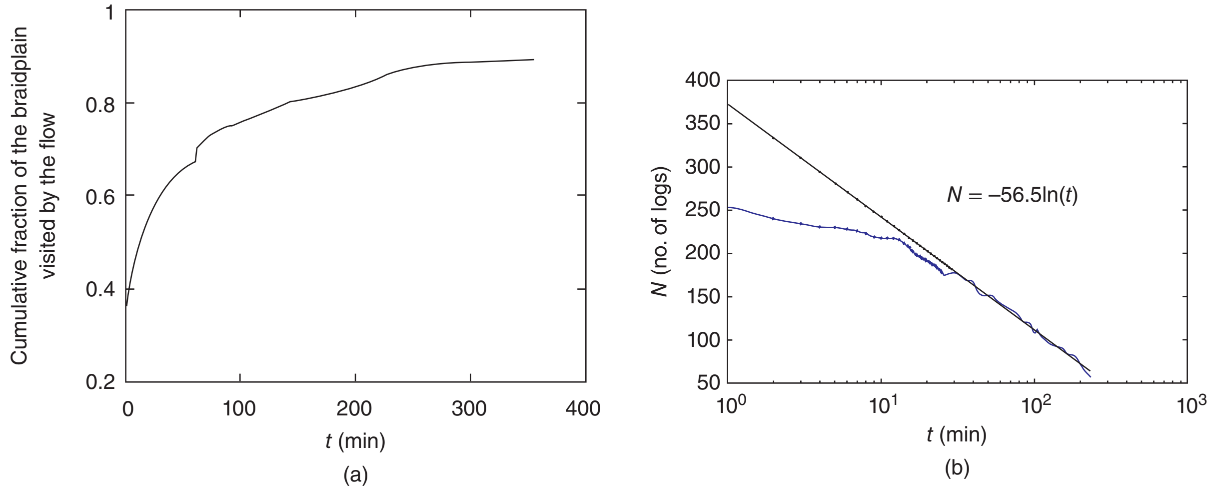


Figure 20.8 (a) Cumulative fraction of the braidplain reworked by the flow for a steady-state braided channel (Tal and Paola 2010). (b). Reduction in log count through time for the experimental braid plain shown in Fig. 20.7. Bottom curve: number of plastic particles simulating logs present on the experimental braidplain. Straight line: linear regression.

required for the entire braid plain to be reworked (cf. Cazanacli *et al.* 2002).

Instream structures

Stream manipulation often involves emplacement of artificial structures in the river channel – even if, in keeping with the spirit of restoring natural characteristics, these are typically constructed of natural materials (stone, wood) rather than concrete. Examples include rock weirs, fish habitats ('lunker structures') and J-hooks. There has been relatively limited controlled study (e.g. Papanicolaou *et al.* 2005) of how these structures work and more importantly how they fail (Radspinner *et al.* 2010). Figure 20.9 shows an example of flow visualization around a rock vane, a form of rock structure used commonly in stream restoration, performed at SAFL. In most cases the flow and sediment-transport fields around such structures have not been studied in detail. The SAFL approach is to couple these measurements to turbulence models (Kang *et al.* 2011) that can resolve the detailed structure of the three-dimensional turbulence that the structures produce.

A recent set of experiments examining flow fields, sediment transport and nutrient dynamics in the vicinity of in-stream rock structures in the OSL demonstrates the capability of full-scale experiments, coupled with indoor, field and numerical investigations, to provide insight into the complex interactions between physical, chemical and biological ecogeomorphic processes. High-resolution topography illustrates the local effect of structures on scour and deposition within the stream channel (Fig. 20.10). In addition to direct measurements of flow field and scour in the vicinity of these structures, this topography can be used as model input and validation for flow (Kang *et al.* 2011) and bed morphodynamics modelling (Khosronejad *et al.* 2011). The modelling addresses the need for structure design using physically based hydraulic engineering principles (Radspinner *et al.* 2010).

Figure 20.9 Surface flow visualization in the OSL using white confetti and long-exposure photographs. Visualizations of this type provide a way of validating computational modelling results and provide insight into the complex flow patterns around instream structures. Flow is from right to left.

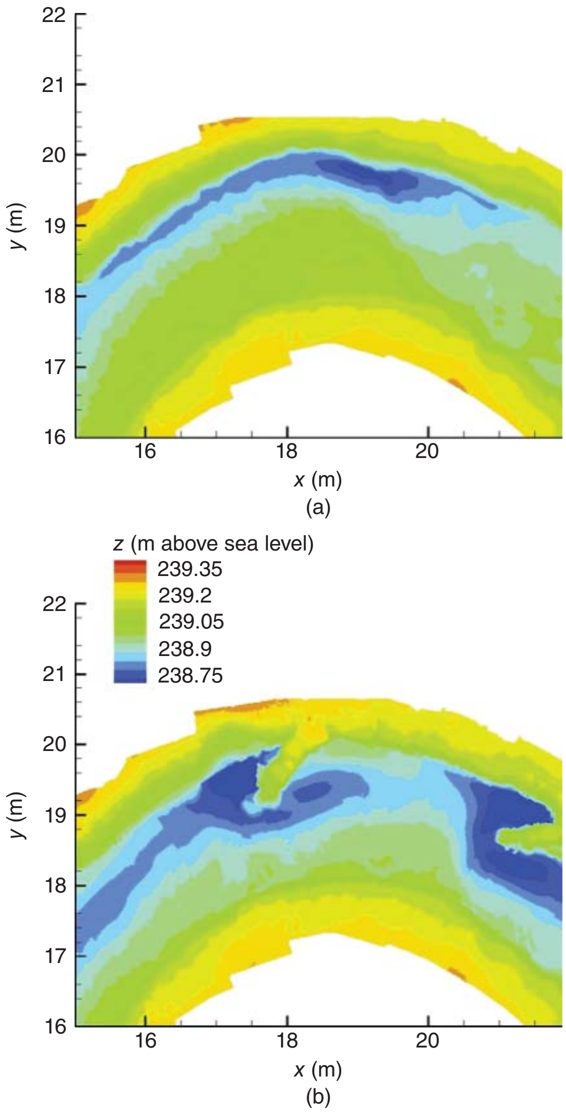


Figure 20.10 Measured topography in the OSL (a) without and (b) with rock vanes, with a mobile sand bed. Flow is from left to right.

Although the presumed ecological benefits of instream structures – in terms of habitat, stream stabilization and nutrient processing – have led to their adoption by many federal, state and local regulatory agencies (Johnson *et al.* 2002), post-installation monitoring of stream restoration structures for ecological benefit is lacking (Bernhardt *et al.* 2005). For example, restoration activities that increase carbon supply, increase contact between the water and benthos and increase connections between streams and adjacent terrestrial environments are generally expected to increase instream nitrogen removal, although the perceived benefits of stream restoration on nitrogen removal have yet to be extensively quantified (Craig *et al.* 2008). Taking this understanding a step further, to develop predictive models of stream physical and ecosystem response to restoration activities requires a fundamental understanding of stream

Figure 20.11 Measuring simultaneous dissolved oxygen, temperature, nitrate and velocity in the vicinity of a rock structure in the SAFL Outdoor StreamLab.

processes at spatial scales ranging from microbial to reach scale (the scale of restoration activities) and beyond. Because both water and sediment feed can be controlled, facilities such as the OSL provide a means to conduct controlled experiments incorporating processes such as interactions between flow, structures and biota that cannot be captured in an indoor flume, permitting experiments to quantify coupled flow and water quality parameters in the vicinity of structures (Fig. 20.11).

Future work

As discussed above, focused research is under way on the interaction of flow and sediment with structures commonly used in restoration such as J-hooks and vanes. However, much more work needs to be done on this topic; as it is, structures such as these are often deployed with a level of understanding of how they work that would never be tolerated for classically built engineering structures. The research must couple detailed studies of the flow and sediment dynamics with guidelines for using the results in practice.

Vegetation, fine sediment and the quest for laboratory-scale meandering

Finding ways to induce experimental streams to self-organize so as to reproduce observed planforms is naturally the goal of any experimentalist. Yet experimental failures can also have important consequences and small-scale experiments can turn out to be most helpful in unravelling problems of major geomorphic significance. The understanding of meandering dynamics is one of these. Despite questions raised about the extent to which meandering rivers are the 'natural' fluvial style in a given region (Walter and Merritts 2008; Métivier and Barrier 2012; see also Chapter 7), they are clearly common in the world today and often (if sometimes inappropriately) used as a reference state for restoration projects (Miller and Ritter 1996; Rosgen 1996; Malakoff 2004; Simon *et al.* 2007; Roper *et al.* 2008). Hence it is noteworthy that it has been, up to now, almost impossible to reproduce self-organized and self-maintaining meandering planforms in laboratory experiments (Murray and Paola 1994; Kleinhans 2010). By contrast, braided rivers are easy to produce experimentally using almost any combination of non-cohesive sediment type/size and flows that are above the threshold for sediment motion. Laboratory experiments on braided rivers have led to major insights into these systems (Ashmore and Parker 1983; Ashmore 1985, 1991; Schumm *et al.* 1987; Paola and Foufoula-Georgiou 2001; Federici and Paola 2003; Métivier and Meunier 2003; Meunier *et al.* 2006). On the other hand, sinuous single-thread channels produced in early experiments (Friedkin 1945) were transient; chute cutoffs and widening eventually led to the formation of multiple threads if the experiments ran long enough and/or the experiments were run under conditions of mild degradation that are not representative of many field cases. The results of these experiments support the idea that braiding is the fundamental instability of unconstrained flow over a non-cohesive bed (Murray and Paola 1994; Paola 2001) and that the key to sustainable meandering is to slow the rate of widening and discourage channel cutoffs. Until recently, reproduction of small-scale meanders in laboratory experiments seemed impossible, so that hydraulic models to study flow and sediment transport in meander bends used moulded, fixed meandering channels (Bhuiyan *et al.* 2010). The curvature of such immobile channels is typically theoretically imposed using a sine-generated curvature model. Although interesting, these models are limited in their application to geomorphological problems because their planform cannot evolve.

Schumm *et al.* (1987) made benchmark advances by defining the conditions necessary for the development of stable meanders, using two types of experiments. In the first type, they progressively added bed load sediment at the inlet of a channel carved into more resistant material. Point bars developed that induced channel erosion and initiation of a sinuous channel. In the second set of experiments, they showed that an initial single-thread straight channel carved into non-cohesive alluvium evolved into a sinuous channel before final evolution into a braided stream – this point has been discussed at length by Federici and Paola (2003) and theoretically by Devauchelle *et al.* (2010). Schumm *et al.* (1987) then managed to stop the evolution towards a braided state by adding fine cohesive sediment at the inlet of the experiments. The deposition of these sediments on the point bars stabilized them and helped maintain a single-thread sinuous channel. They therefore suggested that bedload, fine cohesive sediments and bank stability were the necessary ingredients needed to maintain the meandering pattern. Note that meandering here refers to relatively low sinuosity cases.

Whereas a simulation of the onset of channel meandering in the laboratory is relatively easy, until recently no researcher had succeeded in developing self-formed, high-amplitude meandering channels in the laboratory. Several recent studies have made significant advances towards a methodology for producing meandering rivers experimentally, each with a unique approach to stabilizing banks. Smith (1998) used a complex cohesive mixture of kaolinite, cornstarch, white China clay and diatomaceous earth. Smith was able to develop self-formed, highly sinuous channels that actively migrated and constructed point bars. However, the channels eventually became stable (i.e. stopped migrating) and did not develop active cutoffs and regeneration of bends, key features of actively meandering rivers. The important conditions for the formation of meanders in Smith's experiments were readily identifiable (Lajeunesse *et al.* 2010b). Of critical importance was his use of easily transported fine-grained sediments with slight cohesion. The small grain size of the sediment, on the order of 4–30 μm, enabled point bar deposits to form readily in Smith's microscale river, the flow depth of which varied typically between 5 and 7 mm. Formative bed slopes ranged from 0.007 to 0.02 and formative flow discharges ranged from 10 to 40 mL s^{-1}. Cohesion encouraged the formation of well-defined, single-thread channels. Cohesion also allowed point bars to consolidate over time so as to be resistant to subsequent erosion. The sediment input was not closely controlled but input manually twice a day. Images of the streams show that terraces formed (figs 1 and 2 in Smith 1998). Therefore, the flux of sediment probably was only approximately balanced and channels were somewhat incisional. It is difficult to know the influence of this incision on the development and evolution of the stream pattern.

About 10 years after Smith's original experiment, Tal and Paola (2007, 2010) used real vegetation to add bank stability to a bed of purely non-cohesive sand. One of the main goals of the experiments was to determine if the addition of plants alone could cause a transition from braiding to single thread. Rather than trying to maintain a single-thread sinuous channel that developed from a straight channel, by preventing widening and splitting, these experiments were the first in which a fully braided channel evolved to a single-thread channel on its own. As such, the initial condition for the experiments was steady-state braiding in non-cohesive sand under uniform discharge. From here, an experiment consisted of repeated cycles alternating a short-duration high flow with a long-duration low flow. Alfalfa (*Medicago sativa*) was used to simulate vegetation; seeds were uniformly dispersed over the bed at the end of each high flow. Plants established on freshly deposited bars and areas of braid plain that were unoccupied during low flow. The presence of the plants had the effect of progressively focusing the high flow so that a single dominant channel developed. The plants produced the two key effects required to develop experimental meandering: slowing the rate of widening and discouraging channel cutoffs. Once a single-thread channel developed, the establishment of new vegetation ensured that deposition along the inner bank was able to keep up with erosion along the outer bank, thus allowing the channel to migrate actively while maintaining roughly constant width. An advantage of allowing the single-thread channel to self-evolve from a braided channel is that it organized itself to a geometry that was just sized to carry the high flow. Sediment was supplied at a constant rate at the upper end of the experiment during high-flow periods. Therefore, the braided channel that was able to carry the imposed load metamorphosed into a new, less efficient channel, as evidenced by the onset of sediment deposition within the experiment after vegetation growth. As emphasized by Métivier and Barrier (2012), although bed load supply is necessary to produce point bars, it must not be too high in order for a stable single-thread plan form to develop.

As proposed by Paola (2001) and demonstrated by Murray and Paola (2003) and Hicks *et al.* (2008), the time-scale for establishment of vegetation relative to a characteristic channel or bed mobility time-scale is a key organizing parameter in river systems influencing channel planform. Similar considerations appear to control plant selection also (Crouzy and Perona 2012; Perona *et al.* 2012). An equilibrium between the space–time characteristics of seed dispersal and plant growth and the occupation, abandonment and reworking of the bed by the flow should eventually be reached, resulting in an equilibrium mean channel width and area permanently occupied by vegetation. In the SAFL experiments described above, the relative time-scales of vegetation growth and channel migration rates were systematically varied by changing the flood frequency (i.e. the time vegetation had to establish between floods) and duration of floods (i.e. the fraction of the river bed that was reworked during a flood). Experiments in which the time-scale of vegetation growth was short relative to channel migration time developed into single-thread channels with a substantially reduced wetted width and the development of an extensive floodplain permanently covered by vegetation. An experiment with high channel migration rates maintained multiple active channels, a higher wetted width and a lower vegetated area. Overall, the vegetation–channel migration time-scale ratio appears to be an important parameter in controlling channel planform and dynamics.

Several researchers have also experimented with the use of light-weight sediment in promoting meander development. In addition to increasing the cohesiveness of the banks, fine material that settles out of suspension builds the height of the point bar and fills in chute-channels, helping to deter the flow from splitting. Peakall *et al.* (2007) reported experiments that produced meandering using a combination of sand and silica flour, however, water discharges in the experiments were kept deliberately low to prevent flow from overtopping point bars and banks and occupying new channels. Braudrick *et al.* (2009) used the methodology developed by Tal and Paola (2007) in combination with light-weight sediment to stabilize a single-thread channel which evolved from a straight channel. An initial meander was imposed at the upstream end of the experiment. This initial pattern induces the formation of two successive curves downstream during the course of the experiment. Although the form is clearly meandering, the amplitude of the initial meander remains higher than that of the two self-formed curves downstream of it, suggesting that the influence of initial conditions remains and the effect is partly that of a forced oscillator. Another approach to creating experimental meanders is being developed at the University of Utrecht, who also use an initial condition to help create the meandering, in this case periodic lateral oscillation of the sediment–water feed point (Dijk *et al.*, 2012). This approach also has elements of the forced oscillator about it, but as with the method of Braudrick *et al.* (2009), it is able to produce single-thread channels with at least moderate sinuosities at laboratory scales.

The practical interest in having a simple and reliable methodology for studying meandering in a laboratory setting where variables can be easily controlled and varied, is evident by the numerous ongoing projects to restore meandering rivers across Europe and North America to a more natural and ecologically healthy state. That numerical models still lack fundamental, physically based methods for important processes such as sediment transport and vegetation dynamics underlines the need for laboratory experiments on meandering to advance the state of the science and guide restoration projects. Yet the set of experiments reported above still leaves questions: is meandering stable? Is it 'fragile' and, if so, under what conditions? What about the conclusion of Schumm (1963) that gravel bed meandering rivers were more or less 'no bedload transport' streams? Answers to these questions have important consequences; for example, if meandering is less robust than generally thought, then imposing it on streams is likely to lead to more unfortunate examples such as the well-known case of Uvas Creek reported by Kondolf *et al.* (see Chapter 7). In the context of this chapter, it is interesting to note that so far *only small-scale experiments* (Smith 1998; Tal and Paola 2007) have come close to success. The difficulty of producing meandering at laboratory scales is in itself a potential source of insight about the nature of meandering that has not been fully exploited to date. Nonetheless, the results summarized above together indicate that a variety of techniques exist that enable experimentalists to reproduce self-organized dynamic meanders at least to moderate sinuosities, at laboratory scales. The main elements include some form of bank stabilization such as vegetation maintained by repeated seeding, and fine and/or light-weight sediment, with possible help from an upstream bend or oscillating supply point. These developments offer hope for applying experimental methods to the management and restoration of meandering single-thread channels.

Other biotic interactions with rivers

The focus on meandering discussed above in riparian vegetation studies arises because of the practical and applied motivation for being able to reproduce meandering rivers experimentally. However, vegetation is important in its own right, mediating flow and nutrient transfer (Nepf 2004; Lightbody and Nepf 2006) and also for its effect on sediment stability (e.g. Rominger *et al.* 2010) and sedimentation patterns (Zong and Nepf 2010). A recent line of experimental research at EPFL (Lausanne) has focused on a different facet of the plant-stream coupling: plant selection by flooding, using an experimental approach in which repeated floods are applied to growing vegetation (Perona *et al.* 2012). The selection process is mediated by the mechanics of plant removal (Edmaier *et al.* 2011). The result of this is a stochastic model of plant selection by floods (Crouzy and Perona 2012) that emphasizes the key role of variance of plant characteristics in the selection process. This approach has great promise for advancing ecological forecasting, especially as the additional effects of stochastic variation in flooding, substrate and morphology are added.

There are many other potential avenues for experimental research on the interplay of physical and biotic processes in rivers. There is a large body of research evaluating fish swimming ability in controlled laboratory settings. We have not included topics such as the effects of flow and turbulence on fish and design of streams for fish habitat, because these do not directly influence stream morphology and they have fairly large and mature literature of their own. We note the recent development of experimental streams to test the effect of stream restoration on fish growth directly . Experimental research in stream biology in some cases requires work at more or less full scales (Wilcock *et al.* 2008) because important aspects of the biota in question are scale specific; this is one of the main motivations for facilities such as the SAFL Outdoor StreamLab. Other new large-scale experimental facilities aimed at riparian ecological processes are being developed that, like the OSL, aim to provide controlled but near-natural conditions, for example, the Baylor Experimental Aquatic Research (BEAR) stream facility (King *et al.* 2012) and the Aquatic Ecology Laboratory at Oak Ridge National Laboratory (http://www.esd.ornl.gov/facilities/aquatic™uscore;ecology™uscore;lab.shtml).

Another important area where biology and physical processes are linked and amenable to experimental study is hyporheic flow, i.e. shallow porous-media flow in and around the channel. For example, Sawyer *et al.* (XXX) demonstrated the effect of channel-spanning logs in a laboratory flume on hyporheic flow and temperature; Nowinski *et al.* (2011) studied the hyporheic transport of fine materials through a meander in the OSL experimental facility; Tonina and Buffington (2007) combined experiments and numerical models to investigate hyporheic flow induced by pool-riffle structure in gravel rivers; and Jin *et al.* (2011) and Kessler *et al.* (2012) have shown that even ripple-scale bedforms induce significant hyporheic flow and thus may affect geochemical processes in streams. The multi-scale model of Stonedahl *et al.* (2010) and its associated experimental data sets nicely illustrate the nature and importance of hyporheic flow in rivers and the potential of studying it experimentally.

Future work

The research summarized above makes it apparent that vegetation–physical interactions are a two-way street in which plants both respond to their physical environment and directly modify it. Numerous field observations have alerted us to the fact that changes to a river's flow and/or sediment regime can drive changes in vegetation cover which in turn drive changes in channel planform. One of the challenges, however, lies in isolating the role of individual variables driving these changes. For example, both reduced flows and vegetation expansion can lead to a decrease in active channel width. Experiments are an excellent way to isolate individual variables.

In the context of the vegetation–channel time-scale ratio discussed above as a control on planform, a wide range of parameters influencing both vegetation and channel time-scales can potentially be changed as part of stream restoration and management schemes. These include flood frequency and magnitude and the establishment of invasive plants that may establish faster than native species. This fact is both concerning and encouraging. Experimental studies isolating particular variables would be especially useful in developing our understanding of how physical–biotic feedbacks work, which at this point is fairly rudimentary, providing a basis for better management of physical and biotic resources.

Beyond vegetation, research on 'stream biophysics' – the interplay of biotic, biogeochemical and physical processes – is diversifying and accelerating rapidly. We see particularly strong growth potential in experimental study of microbial and other small-scale biotic processes and they interact with fluid and sediment mechanics, especially for fine sediments.

20.4 Scaling issues and application of experimental results

One of the major stumbling blocks for application of experimental methods in stream geomorphology generally has been the perception that scaling up experimental observations to field conditions is difficult or perhaps impossible. Paola *et al.* (2009) present an extensive analysis of issues related to scaling experiments in geomorphology generally. Paola *et al.* (2009) draw heavily on a paper by Malverti *et al.* (2008) that showed the extent to which experiments with laminar flow reproduce the geometry and behaviours of a wide variety of geomorphic systems. Here we summarize some of the results of this work and how they apply to stream management.

The paper by Malverti *et al.* (2008) documents the remarkable extent to which experimental studies of morphodynamics at scales so small that the flow is laminar reproduce the geometry and behaviour of field cases. This finding is important to the applicability of morphodynamics experiments generally to field conditions because, of all the effects of reproducing streams at laboratory scales, the change from turbulent to laminar flow would seem to be among the most profound and many experimenters have set their systems up to avoid crossing this apparently important threshold. Yet Malverti *et al.* showed that, like the Wizard of Oz, the threat behind the screen is not so terrible after all. Turbidity currents are perhaps the most dramatic illustration of the surprising insensitivity of basic morphodynamics to turbulence: since they rely on sediment suspension for their very existence, one might expect that they would not survive the transition to laminar flow – and yet they do. Malverti *et al.* (2008) explained the insensitivity of morphodynamics of many kinds in terms of the similarity of drag and sediment transport laws between laminar and turbulent flows.

Paola *et al.* (2009) suggest that the findings of Malverti *et al.* (2008) fit a general pattern of what they refer to as 'unreasonable effectiveness' of laboratory experiments in geomorphology and stratigraphy. The unreasonableness arises from the fact that it is effectively impossible to scale typical morphodynamics experiments to the field using classical methods of scaling based on matching dimensionless numbers; the disparity in Reynolds numbers associated with laminar versus turbulent flow is one major example of this. One might simplistically conclude that laboratory experiments should be useless. But classical engineering scaling was developed to provide a means by which an experimental model could be constructed so as to reproduce, under simple algebraic transformations, the distribution of flow and stresses for a field prototype. This was (and is) essential if the laboratory study is to be used as the sole means by which these properties are going to be estimated before a project is built in the field. Even so, for reasons summarized by Paola *et al.*, even for fixed-geometry models the full requirements for dynamic scaling (matching Froude and Reynolds numbers) can almost never be met unless the scale is 1:1. Introducing a typical range of sediment types brings new dimensionless numbers that are even more intractable for full classical scale modelling. Of these, the most important and obvious limitation is that scaling sand in the field down to laboratory scales would typically lead to laboratory grain sizes that are in the cohesive range – a qualitative transition whose consequences are not well understood but are likely to be important for morphodynamics.

However, as Malverti *et al.* (2008) and Paola *et al.* (2009) make clear, laboratory experiments seem to work in the sense that a wide range of natural morphodynamic patterns develop spontaneously at laboratory scales and seem to exhibit many of the same behaviours that they do in the field. This includes the main river channel patterns, especially now that the work discussed above has shown the way to creating self-maintaining meanders. Paola *et al.* (2009) point out that the traditional engineering work-around for the impossibility of matching Reynolds numbers between field and laboratory has been to invoke the empirical observation of Reynolds number (*Re*) independence. The idea is that as long as *Re* is high enough, its precise value is not important for many system-scale phenomena, including in particular the overall flow pattern. The independence of drag coefficient from *Re* over several decades' variation in *Re* is an example of *Re* independence. Paola *et al.* (2009) suggest that there is no reason why this kind of scale independence should be restricted only to the Reynolds number. At least two lines of reasoning including the widespread presence of fractals in landscape patterns and the persistence of morphodynamics in laminar flow pointed out by Malverti *et al.* (2008) suggest that scale independence is common in landscape dynamics generally. Changing our conceptual basis from classical engineering scaling, with its unreachable requirements for insuring commensurability across scales, to widespread scale independence greatly broadens the scope for using experiments in stream science and practice. On the other hand, the fundamental basis for scale independence is much less well understood than the principles of classical scaling. In particular, there is a good deal of work still to be done on the origins and limits of scale independence. In fluvial geomorphology generally, it is best to view laboratory experiments simply as small systems in their own right, whose relevance to the field lies in the similarity of the important dynamics across scales. Because the limits to this similarity are in general not currently known, experiments cannot be treated simply as miniature analogues of field systems. But a clear mechanistic understanding of the observations of interest should carry with it an understanding of how the observations might be expected to vary with scale. Theoretical models are one powerful way of encapsulating our mechanistic understanding, and testing theoretical models remains one of the best uses of morphodynamics experiments generally. Overall, the work summarized in this section should calm the fears of practitioners who worry that results derived from the laboratory do not apply to field scales.

20.5 Additional areas for experimentation

Better mechanistic support for using biota in stream management

Vegetation is often used in stream management but presents far more complex design challenges than traditional hard engineering structures. For example, plants in water can respond adaptively according to sedimentology (encroachment to the bed) and flow dynamics. Bornette and Puijalon (2011) showed that aquatic macrophytes can develop alternative traits to adapt to flow conditions. Avoidance corresponds to minimization of the drag force exerted by the flow (through leaf orientation, for example). Tolerance corresponds to maximization of resistance to breakage (through growth of stem width, for example). Analysis of experiments and phylogeny suggests that these two adaptations are negatively correlated such that there is a trade-off between the two strategies. Flume studies such as that by Bornette and Puijalon (2011) hold great promise in understanding this trade-off, by allowing the establishment of a compendium of resistance/avoidance thresholds for macrophytes to given flow conditions. These in turn can be used to understand what plant families can be expected to develop under given hydrographs and expected typical (bankfull) flow velocities in a restoration project.

More generally, it seems to us that controlled experiments, even if they have to be performed at full scale, would be useful in providing a better mechanistic basis for understanding the stability and best use of bioengineering materials such as fibre bundles and root wads. As discussed earlier, such studies are likely to be most effective if coupled to theoretical analysis, numerical and/or analytical.

Eco-hydrology and river morphology

Stream restoration and experimentation face challenging problems where live organisms are concerned. Among the many problems that researchers have to deal with, one of particular importance is the high sensitivity of animals to boundary conditions and especially to reduction in free space. Hence animals feel the presence of walls and it may affect their reactions to imposed stimuli. In this case, scaling issues must be addressed and replications must be performed in order to ensure reproducibility of the results. For example, Jonsson *et al.* (2006) compared the hydrodynamic performance of 12 flumes used to study benthic organisms. They concentrated on boundary layer dynamics and turbulence because benthic organisms are thought to be highly responsive to these parameters. They showed how most of the classic straight or racetrack flumes provide the necessary conditions to study benthic ecology. They also showed that the size of the organisms that can be studied depends on the height of the turbulent boundary layer, hence potentially on the length of the flume. Therefore, although most geomorphological stream experimentations are only Froude scaled, their turbulence characteristics allow them to be used to address stream restoration questions.

A common goal in stream management is to maintain or increase biodiversity. The importance of ecological diversity implies knowledge of how to maintain and restore diversity through channel and floodplain restoration processes. Bornette and Puijalon (2011), for example, reviewed the abiotic factors affecting macrophyte development: light, temperature, water nutrient content (dissolved CO_2 and HCO_3^-, which are influenced by global atmospheric CO_2 changes, P and N). Species richness is expected to be the highest where nutrient levels are intermediate. When high or low levels are encountered, species that are stress tolerant or strong competitors win out. Flow and the morphological structure of floodplains induce nutrient dispersal and concentrations that vary from the main thread relative to, for example, bars, groundwater and oxbow lakes. Flow paths and patterns in a complex floodplain are therefore key features of a restoration project.

Hyporheic flow

Most experimental river research to date has focused on surface water flow and its interaction with sediment and/or vegetation. However,, as indicated in the previous section, groundwater–surface water interaction is crucial for understanding (at least) nutrient dynamics. Hyporheic flow has received far less attention from the experimental community than it deserves. There is great potential for experiments across a range of scales, to study groundwater flow patterns in bedforms, point bars and floodplains, as exemplified by the experimental and theoretical studies of hyporheic flow cited above

Scale independence and scaling

As discussed above, we believe that to exploit fully the potential of experiments in stream science generally, we must complement the methods of classical scaling with a much better understanding of scale independence in morphodynamics: its origin, mechanism, range and limitations. This can best be done through a combination of theoretical study, especially analysis of sources of scale dependence in the governing equations of morphodynamics, and systematic comparison of systems of varying scale, with a focus on isolating differences directly associated with scale – no small task – all with the aim of replacing 'does this experiment scale?' with a new set of questions along the lines of 'which aspects of this experiment should apply to this field setting, which should not, and why?'.

Microbial processes

At the fine end of the scale range, investigation of microbial processes has accelerated, as the critical role of microbes in mediating geochemical reactions in the surface environment becomes apparent. We believe that there is tremendous scope to develop this area experimentally. Periphyton, a critical element of the riverine food web, can be grown under experimental conditions but is sensitive to light and other aspects of the local environment: for example, see the work of Orr *et al.* (2009) evaluating the impact of geomorphology and periphyton growth on nutrient uptake in a field-scale flume or that of Hondzo and Wang (2002) on the influence of turbulence on periphyton growth. To some extent, the small size of microbes means that the size limitations of experimental systems matter less: for example, microbial biofilms, which influence both physical and geochemical processes, can be grown in experimental flow cells (Zhang *et al.* 2011) and 'microcosms' (Singer *et al.* 2006, 2010), and laboratory setups have been used to good effect to study biogeochemical processes such as metal uptake by aquatic plants and epiphytes (Hansen *et al.* 2011). On the other hand, many types of microbes cannot be grown under laboratory conditions and often potential interactions among microbes and other aspects of the environment, including other microbes and the rest of the ecosystem, are so poorly understood that it is difficult to know to what extent microbial behaviour in the laboratory is representative of field conditions. Nevertheless, the potential importance of microbial biofilms and mats for the dynamics of fine sediments and the contaminants they carry would alone be a worthy subject for a major research effort.

20.6 Conclusion

Practical stream management presents a complex, fascinating set of problems that link hydraulics, geomorphology, ecology and social dynamics. Given the amount being spent on it worldwide, rapid improvement in the present weak scientific basis for stream management is essential. Experimental research to date has been an underused tool for accomplishing this. Laboratory experiments will be most effective when they are carried out not in isolation but as part of a web of field, laboratory and theoretical research. The approaches developed at IPGP and SAFL exemplify this, in somewhat different ways, as discussed in this chapter. In the absence of simple scaling relations, the way forward seems to us to be to combine classical laboratory experiments over varying scales, including outdoor facilities with field work (e.g. instrumented watersheds) and mechanistic theory. This is the best way forward in providing a solid scientific basis for stream sustainable management.

A major obstacle to the growth of experimental studies in support of stream management is the lack of a coherent research programme for improving methods for river management and restoration in which experimental studies would take their natural place. Also, most individual fluvial projects are too small to include experimental studies, especially if, as is often the case, the experiments are aimed at general questions rather than improving the design of that particular project. There is no immediate solution to this, but it seems to us that it would make sense for the various groups, public and private, to join forces to encourage research programmes focused on improving the scientific basis for river management and restoration that could benefit the whole community.

In the meantime, stream practitioners could consider the following steps:

1. Contact local universities to see what experimental facilities are available.
2. Consider alternative funding mechanisms; for example, the Internet should make it possible for many small actors to pool modest amounts of money to support targeted research on key stream dynamic questions. This mode would be especially suitable to experimental research.
3. Stream management has a number of regional advocacy organizations such as PRRSUM (http://www.prrsum.org/) that could provide a means of organizing and collaboratively funding basic research relevant to stream management.
4. In the past, US government agencies have developed interagency groups to work on specific cross-agency research problems. Perhaps the time has come to organize similar efforts, on an international level, for the experimental study of fundamental questions for practical stream management such as those identified in this chapter.

Acknowledgements

C.P. and J.L.K. acknowledge with gratitude the contributions of Dr Anne Lightbody and the SAFL undergraduate interns to their research, and also financial support from the NSF via the National Center for Earth-surface Dynamics (NCED), Agreement Number EAR-0120914), the National Cooperative Highway Research Program (NCHRP) Project No. 24-33, and a collaborative agreement with Yonsei University (Seoul). This publication was partially funded by ANR GESTRANS. Comments and suggestions from C. Braudrick, an anonymous reviewer and the Editors are gratefully acknowledged. This is IPGP contribution No. 3715.

References

Abbe, T., Montgomery, D.R. 1996. Large woody debris jams, channel hydraulics and habitat formation in large rivers. *Regulated Rivers – Research and Management* 12: 201–221.

Amoros, C. and Bornette, G. 2002. Connectivity and biocomplexity in waterbodies of riverine floodplains. *Freshwater Biology* 47(4): 761–776.

Andrews, E.D. 1984. Bed-material entrainment and hydraulic geometry of gravel-bed rivers in Colorado. *Geological Society of America Bulletin* 95: 371–378.

Armstrong, L. 2003. *Bank Erosion and Sediment Transport in a Microscale Straight River*, PhD Thesis, Université Paris 7 – Denis Diderot.

Ashmore, P.E. 1985. *Process and Form in Gravel Braided Streams: Laboratory Modelling and Field Observations*, PhD Thesis, University of Alberta, 414pp.

Ashmore, P.E. 1991. How do gravel-bed rivers braid? *Canadian Journal of Earth Sciences* 28: 326–341.

Ashmore, P. and Parker, G. 1983. Confluence scour in coarse braided streams. *Water Resources Research* 19(2): 392–402.

Ashworth, P.J., Best, J.L., Leddy, J.O. and Geehan, G.W. 1994. The physical modelling of braided rivers and deposition of fine-grained sediment. In: Kirkby, J.M., ed., *Process Models and Theoretical Geomorphology*, Chichester: John Wiley & Sons, pp. 115–139.

Bennett, S.J. and Simon, A., eds. 2004. *Riparian Vegetation and Fluvial Geomorphology, Water Science and Application Series, vol. 8*, Washington, DC: American Geophysical Union, 282pp.

Bennett, S.J., Pirim, T. and Barkdoll, B.D. 2002. Using simulated emergent vegetation to alter stream flow direction within a straight experimental channel. *Geomorphology* 44(1–2): 115–126.

Bennett, S.J., Wu, W., Alonso, C.V. and Wang, S.S.Y. 2008. Modeling fluvial response to in-stream woody vegetation: implications for stream corridor restoration. *Earth Surface Processes and Landforms* 33(6): 890–909.

Bernhardt, E.S., Palmer, M.A., Allan, J.D., Alexander, G., Barnas, K., Brooks, S., et al. 2005. Synthesizing U.S. river restoration efforts. *Science* 308(5722): 636–637.

Bhuiyan, F., Hey, R.D. and Wormleaton, P.R. 2010. Bank-attached vanes for bank erosion control and restoration of river meanders. *Journal of Hydraulic Engineering* 136(9): 583–596.

Bocchiola, D., Rulli, M.C. and Rosso, R. 2006. Flume experiments on wood entrainment in rivers. *Advances in Water Resources* 29(8): 1182–1195.

Bonnet, S. and Crave, A. 2003. Landscape response to climate change; insights from experimental modeling and implications for tectonic versus climatic uplift of topography. *Geology* 31(2): 123–126.

Bonnet, S. and Crave, A. 2006. Macroscale dynamics of experimental landscapes. In: Buiter, S.J.H. and Schreurs, G., eds., *Analogue and Numerical Modelling of Crustal-scale Processes, Special Publication No. 253*, London: Geological Society, pp. 327–339.

Bornette, G. and Puijalon, S. 2011. Response of aquatic plants to abiotic factors: a review. *Aquatic Sciences* 73: 1–14.

Braudrick, C. and Grant, G.E. 2000. When do logs move in rivers? *Water Resources Research* 36(2): 571–584.

Braudrick, C.A. and Grant, G.E. 2001. Transport and deposition of large woody debris in streams: a flume experiment. *Geomorphology* 41: 263–284.

Braudrick, C.A., Dietrich, W.E., Leverich, G.T. and Sklar, L.S. 2009. Experimental evidence for the conditions necessary to sustain meandering in coarse-bedded rivers. *Proceedings of the National Academy of Sciences of the United States of America* 106: 16936–16941.

Cazanacli, D., Paola, C. and Parker, G. 2002. Experimental steep, braided flow: application to flooding risk on fans. *Journal of Hydraulic Engineering* 128(3): 322–330.

Coleman, S.E. and Eling, B. 2000. Sand wavelets in laminar open-channel flows. *Journal of Hydraulic Research* 38(5): 331–338.

Craig, L.S., Palmer, M.A., Richardson, D.C., Filoso, S., Bernhardt, E.S., Bledsoe, B.P., et al. 2008. Stream restoration strategies for reducing river nitrogen loads. *Frontiers in Ecology and the Environment* 6(10): 529–538.

Crouzy, B. and Perona, P. 2012. Biomass selection by floods and related timescales. Part 2. Stochastic modeling. *Advances in Water Research* 39: 97–105.

Daubrée, A. 1879. *Études Synthétiques de Géologie Expérimentale*, Paris: Dunod.

Davies, N.S. and Gibling, M.R. 2010. Cambrian to Devonian evolution of alluvial systems: the sedimentological impact of the earliest land plants. *Earth-Science Reviews* 98(3–4): 171–200.

Devauchelle, O., Malverti, L., Lajeunesse, E., Josserand, C., Lagree, P.Y. and Métivier, F. 2010. Rhomboid beach pattern: a laboratory investigation. *Journal of Geophysical Research, Earth Surface* 115: F02017.

Dijk, W., Lageweg, W., and Kleinhans, M.: Experimental meandering river with chute cutoffs, Journal of Geophysical Research: Earth Surface (2003–2012), 117, 2012.

Edmaier, K., Burlando, P. and Perona, P. 2011. Mechanisms of vegetation uprooting by flow in alluvial non-cohesive sediment. *Hydrology and Earth System Sciences* 15: 1615–1627.

Federici, B. and Paola, C. 2003. Dynamics of channel bifurcation in non-cohesive sediments. *Water Resources Research* 39(6): 3-1–3-15.

Friedkin, J.F. 1945. *A Laboratory Study of the Meandering of Alluvial Rivers*, Vicksburg, MS: US Army Corps of Engineers Waterways Experiment Station.

Gurnell, A.M. 2003. Wood storage and mobility In: Gregory, S., Boyer, K. and Gurnell, A., eds., *The Ecology and Management of Wood in World Rivers, Symposium 37*, Bethesda, MD: American Fisheries Society, pp. 75–91.

Gurnell, A.M., Piegay, H., Swanson, F.J. and Gregory, S.V. 2002. Large wood and fluvial processes. *Freshwater Biology* 47: 601–619.

Hansen, A., Stark R. and Hondzo M. 2011. Uptake of dissolved nickel by Elodea canadensis and epiphytes influenced by fluid flow conditions, *Hydrobiologia* 658(1): 127–138.

Hession, W.C., Pizzuto, J.E., Johnson, T.E. and Horwitz, R.J. 2003. Influence of bank vegetation on channel morphology in rural and urban watersheds. *Geology* 31(2): 147–150.

Hey, R.D. and Thorne, C.R. 1986. Stable channels with mobile gravel beds. *Journal of Hydraulic Engineering* 112(8): 671–689.

Hicks, D.M., Duncan, M.J., Lane, S.N., Tal, M. and Westaway, R. 2008. Contemporary morphological change in braided gravel-bed rivers: new developments from field and laboratory studies, with particular reference to the influence of riparian vegetation. In: Habersack, H., Piégay, H. and Rinaldi, M., eds., *Gravel-bed Rivers VI: from Process Understanding to River Restoration*, Amsterdam: Elsevier, pp. 557–586.

Hondzo, M. and H. Wang (2002), Effects of turbulence on growth and metabolism of periphyton in a laboratory flume, *Water Resources Research* 38(12): 13.11–13.19.

Huang, H.Q. and Nanson, G.C. 1997. Vegetation and channel variation; a case study of four small streams in southeastern Australia. *Geomorphology* 18(3–4): 237–249.

Jaggar, T.A. 1908. Experiments illustrating erosion and sedimentation. *Bulletin of the Museum of Comparative Zoology at Harvard College* 49(Geological Series 8): 283–304.

Jiang, Z., Perona, P., Francis, R., Molnar, P. and Burlando, P. 2009. An experimental comparison of silica gel and quartz sand grains as sediment media for growing vegetation at the laboratory scale. *Aquatic Sciences* 71(3): 350–355.

Jin, G., Tang, H., Li, L. and Barry, D.A. 2011. Hyporheic flow under periodic bed forms influenced by low-density gradients. *Geophysical Research Letters* 38(22): L22401.

Johnson, P.A., Tereska, R.L. and Brown, E.R. 2002. Using technical adaptive management to improve design guidelines for urban instream structures. *Journal of the American Water Resources Association* 38(4): 1143–1152.

Jonsson, P.R., Van Duren, L.A., Amielh, M., Asmus, R., Aspden, R.J., Daunys, D., et al. 2006. Making water flow: a comparison of the hydrodynamic characteristics of 12 different benthic biological flumes. *Aquatic Ecology* 40(4): 409–438.

Kang, S., Lightbody, A., Hill, C. and Sotiropoulos, F. 2011. High-resolution numerical simulation of turbulence in natural waterways. *Advances in Water Resources* 34(1): 98–113.

Keller, E.A. and Swanson, F.J. 1979. Effects of large organic material on channel form and fluvial processes. *Earth Surface Processes and Landforms* 4: 361–380.

Kessler, A.J., Glud, R.N., Cardenas, M.B., Larsen, M., Bourke, M. and Cook, P.L.M. 2012. Quantifying denitrification in rippled permeable sands through combined flume experiments and modelling. *Limnology and Oceanography* 57(4): 1217–1232.

Khosronejad, A., Kang, S., Borazjani, I. and Sotiropoulos, F. 2011. Curvilinear immersed boundary method for simulating coupled flow and bed morphodynamic interactions due to sediment transport phenomena. *Advances in Water Resources* 34(7): 829–843.

King, R.S., Walker, C.M., Whigham, D.F., Baird, S.J. and Back, J.A. 2012. Catchment topography and wetland geomorphology drive macroinvertebrate community structure and juvenile salmonid distributions in south-central Alaska headwater streams. *Freshwater Science* 31(2): 341–364.

Kleinhans, M.G. 2010. Sorting out river channel patterns. *Progress in Physical Geography* 34(3): 287–326.

Lague, D., Crave A. and Davy, P. 2003. Laboratory experiments simulating the geomorphic response to tectonic uplift. *Journal of Geophysical Research, Solid Earth* 108(B1): ETG3-1–ETG3-20.

Lai, S.Y.J. and Capart, H. 2009. Reservoir infill by hyperpycnal deltas over bedrock. *Geophysical Research Letters* 36: L08402.

Lajeunesse, E., Malverti, L. and Charru, F. 2010a. Bedload transport in turbulent flow at the grain scale: experiments and modeling. *Journal of Geophysical Research* 115: F04001.

Lajeunesse, E., Malverti, L., Lancien, P., Armstrong, L., Métivier, F., Coleman, S., Smith, C.E., Davies, T., Cantelli, A. and Parker, G. 2010b. Fluvial and submarine morphodynamics of laminar and near-laminar flows: a synthesis. *Sedimentology* 57: 1–26.

Lancien, P., Métivier, F. and Lajeunesse, E. 2005. Simulating submarine channels in flume experiments. *Physicalia Magazine* 27: 377–384.

Lightbody, A.F. and Nepf, H.M. 2006. Prediction of velocity profiles and longitudinal dispersion in emergent salt marsh vegetation. *Limnology and Oceanography* 51(1): 218–228.

Limare, A., Tal, M., Reitz, M., Lajeunesse, E. and Métivier, F. 2011. Optical method for measuring bed topography and flow depth in an experimental flume. *Solid Earth* 2: 143–154.

Lisle, T.E. 1995. Effects of coarse woody debris and its removal on a channel affected by the 1980 eruption of Mount St. Helens, Washington. *Water Resources Research* 31(7): 1797–1808.

Malakoff, D. 2004. Dave Rosgen: the river doctor. *Science* 305(5686): 937–939.

Malverti, L., Lajeunesse, E. and Métivier, F. 2008. Small is beautiful: upscaling microscale experimental results to the size of natural rivers. *Journal of Geophysical Research* 113: F04004.

Manga, M. and Kirchner, J.W. 2000. Stress partitioning in streams by large woody debris. *Water Resources Research* 36(8): 2373–2379.

Merten, E.C., Hintz, W.D., Lightbody, A.F. and Wellnitz, T. 2010. Macroinvertebrate grazers, current velocity and bedload transport rate influence periphytic accrual in a field-scale experimental stream. *Hydrobiologia* 652(1): 179–184.

Métivier, F. and Barrier, L. 2012. Alluvial landscape evolution: what do we know about metamorphosis of gravel bed meandering and braided streams? In: Church, M., Biron, P.M. and Roy, A.G., eds., *Gravel-bed Rivers: Processes, Tools, Environments*, Chichester: John Wiley & Sons, pp. 474–501.

Métivier, F. and Meunier, P. 2003. Input and output mass flux correlations in an experimental braided stream: implications on the dynamics of bed load transport. *Journal of Hydrology* 271: 22–38.

Métivier, F., Lajeunesse, E. and Cacas, M.C. 2005. Submarine canyons in the bathtub. *Journal of Sedimentary Research* 75: 6–11.

Meunier, P., Métivier, F., Lajeunesse, E., Mériaux, A.S. and Faure, J. 2006. Flow pattern and sediment transport in a braided river: the 'torrent de St Pierre' (French Alps). *Journal of Hydrology* 330(3–4): 496–505.

Miller, J. and Ritter, J. 1996. An examination of the Rosgen classification of natural rivers. *Catena* 27: 295–299.

Montogmery, D.R., Collins, B.D. and Buffington, J.M. 2003. Geomorphic effects of wood in rivers. In: Gregory, S., Boyer, K. and Gurnell, A., eds., *The Ecology and Management of Wood in World Rivers, Symposium 37*, Bethesda, MD: American Fisheries Society. pp. 21–47.

Moulin, B. and Piégay, H. 2004. Characteristics and temporal variability of large woody debris trapped in a reservoir on the River Rhone: implications for river basin management. *River Research and Applications* 20: 79–97.

Murray, A.B. and Paola, C. 1994. A cellular model of braided rivers. *Nature* 371: 54–57.

Murray, A.B. and Paola, C. 2003. Modelling the effect of vegetation on channel pattern in bedload rivers. *Earth Surface Processes and Landforms* 28: 131–143.

Nakamura, F. and Swanson, F.J. 1993. Effects of coarse woody debris on morphology and sediment storage of a mountain stream system in western Oregon. *Earth Surface Processes and Landforms* 18: 43–61.

Nepf, H.M. 2004. Vegetated flow dynamics. In: Fagherazzi, S., Marani, M. and Blum, L.K., eds., *Ecogeomorphology of Tidal Marshes, Coastal and Estuarine Studies*, vol. 59, Washington, DC: American Geophysical Union, pp. 137–163.

Nowinski, J.D., Cardenas, M.B. and Lightbody, A.F. 2011. Evolution of hydraulic conductivity in the floodplain of a meandering river due to hyporheic transport of fine materials. *Geophysical Research Letters* 38(1): L01401.

Orr, C.H., Clark, J.J., Wilcock, P.R., Finlay, J.C. and Doyle, M.W. 2009. Comparison of morphological and biological control of exchange with transient storage zones in a field-scale flume. *Journal of Geophysical Research* 114(G2): G02019.

Paola, C. 2001. Modelling stream braiding over a range of scales. In: Mosley, M.P., ed., *Gravel-bed Rivers V*, Wellington: New Zealand Hydrological Society, pp. 11–38.

Paola, C. and Foufoula-Georgiou, E. 2001. Statistical geometry and dynamics of braided rivers. In: Mosley, M.P., ed., *Gravel-bed Rivers V*, Wellington: New Zealand Hydrological Society, pp. 47–71.

Paola, C., Straub, K., Mohrig, D.C. and Reinhardt, L. 2009. The 'unreasonable effectiveness' of stratigraphic and geomorphic experiments. *Earth-Science Reviews* 97: 1–43.

Papanicolaou, A.N., Hobbs, B., Kramer, C., Fox, J.F. and Kjos, L. 2005. Fluid–sediment dynamics around a barb: an experimental case study of a hydraulic structure for the Pacific Northwest. *Canadian Journal of Civil Engineering* 32(5): 853–867.

Peakall, J., Ashworth, P.J. and Best, J.L. 2007. Meander-bend evolution, alluvial architecture and the role of cohesion in sinuous river channels: a flume study. *Journal of Sedimentary Research* 77: 197–212.

Perona, P., Molnar, P., Crouzy, B., Perucca, E., Jiang, Z., McLelland, S., Wüthrich, D., Edmaier, K., Francis, R., Camporeale, C. and Gurnell, A. 2012. Biomass selection by floods and related timescales. Part 1: experimental observations. *Advances in Water Research* 39: 85–96.

Radspinner, R.R., Diplas, P., Lightbody, A.F. and Sotiropoulos, F. 2010. River training and ecological enhancement potential using in-stream structures. *Journal of Hydraulic Engineering* 136(12): 967–980.

Reinhardt, L., Jerolmack, D., Cardinale, B.J., Vanacker, V. and Wright, J. 2010. Dynamic interactions of life and its landsape: feedbacks at the interface of geomorphology and ecology. *Earth Surface Processes and Landforms* 35: 78–101.

Reitz, M.D., Jerolmack, D.J. and Swenson, J.B. 2010. Flooding and flow path selection on alluvial fans and deltas. *Geophysical Research Letters* 37(6): L06401.

Rice, S.P., Lancaster, J. and Kemp, P. 2010. Experimentation at the interface of fluvial geomorphology, stream ecology and hydraulic engineering and the development of an effective, interdisciplinary river science. *Earth Surface Processes and Landforms* 35(1): 64–77.

Rominger, J.T., Lightbody, A.F. and Nepf, H.M. 2010. Effects of added vegetation on sand bar stability and stream hydrodynamics. *Journal of Hydrauli Engineering* 136(12): 994–1002.

Roper, B.B., Buffington, J.M., Archer, E., Moyer, C. and Ward, M. 2008. The role of observer variation in determining Rosgen stream types in northeastern Oregon mountain streams. *Journal of the American Water Resources Association* 44(2): 417–427.

Rosgen, D. 1996. *Applied River Morphology*, Pagosa Springs, CO: Wildland Hydrology.

Schumm, S.A. 1963. *A Tentative Classification of Alluvial River Channels, Geological Survey Circular 477*, Washington, DC: US Geological Survey.

Schumm, S.A. 1977. *The Fluvial System*, New York: John Wiley & Sons, 338pp.

Schumm, S.A., Mosley, M.P. and Weaver, W.E. 1987. *Experimental Fluvial Geomorphology*, New York: John Wiley & Sons, 428pp.

Simon, A., Doyle, M., Kondolf, M., Shields, F.D., Rhoads, B. and McPhillips, M. 2007. Critical evaluation of how the Rosgen classification and associated 'Natural Channel Design' methods fail to integrate and quantify fluvial processes and channel response1. *Journal of the American Water Resources Association* 43(5): 1–15.

Singer, G., Besemer, K., Hödl, I., Chlup, A., Hochedlinger, G., Stadler, P. and Battin, T.J. 2006. Microcosm design and evaluation to study stream microbial biofilms. *Limnology and Oceanography: Methods* 4: 436–447.

Singer, G., Besemer, K., Schmitt-Kopplin, P., Hödl, I. and Battin, T.J. 2010. Physical heterogeneity increases biofilm resource use and its molecular diversity in stream mesocosms. *PLoS ONE* 5(4): e9988.

Smith, C.E. 1998. Modeling high-sinuousity meanders in a small flume. *Geomorphology* 25: 19–30.

Stonedahl, S.H., Harvey J.W., Wörman, A., Salehin, M. and Packman, A.I. 2010. A multiscale model for integrating hyporheic exchange from ripples to meanders. *Water Resources Research* 46(12): W12539.

Tal, M. and Paola, C. 2007. Dynamic single-thread channels maintained by the interaction of flow and vegetation. *Geology* 35(4): 347–350.

Tal, M. and Paola, C. 2010. Effects of vegetation on channel morphodynamics: results and insights from laboratory experiments. *Earth Surface Processes and Landforms* 35(9): 1014–1028.

Tal, M., Frey, P., Kim, W., Lajeunesse, E., Limare, A. and Métiver, F. 2012. The use of imagery in laboratory experiments. In: Carbonneau P. and Piégay H., eds., *Fluvial Remote Sensing for Science and Management*, Chichester: John Wiley & Sons, pp. 299–321.

Tonina, D. and Buffington, J.M. 2007. Hyporheic exchange in gravel bed rivers with pool-riffle morphology: laboratory experiments and three-dimensional modeling. *Water Resources Research* 43(1): W01421.

Walter, R.C. and Merritts, D.J. 2008. Natural streams and the legacy of water-powered mills. *Science* 319(5861): 299–304.

Wilcock, P.R., Orr, C.H. and Marr, J.D.G. 2008. The need for full-scale experiments in river science. *Eos Transactions American Geophysical Union* 89(1): 6.

Wilcox, A.C. and Wohl, E.E. 2006. Flow resistance dynamics in step-pool stream channels: 1. Large woody debris and controls on total resistance. *Water Resources Research* 42(5): W05418.

Yalin, M.S. 1971. *Theory of Hydraulic Models*, London: Macmillan.

Zhang, W., Sileika, T.S., Chen, C., Liu, Y., Lee, J. and Packman, A.I. 2011. A novel planar flow cell for studies of biofilm heterogeneity and flow–biofilm interactions. *Biotechnology and Bioengineering* 108(11): 2571–2582.

Zimmerman, R.C., Goodlett, J.C. and Comer, G.H. 1967. The influence of vegetation on channel form of small streams. In: *Symposium on River Morphology, Publication 75*, Wallingford: International Association of Scientific Hydrology, pp. 255–275.

Zong, L. and Nepf, H. 2010. Flow and deposition in and around a finite patch of vegetation. *Geomorphology* 116(3–4): 363–372.

CHAPTER 21

Statistics and fluvial geomorphology

Hervé Piégay and Lise Vaudor
Université de Lyon, UMR 5600 CNRS, Lyon, France

21.1 Introduction

Current and future use of statistical tools by fluvial geomorphologists

Why a chapter about statistics in a book on fluvial geomorphology? Over recent decades, geomorphologists have applied the laws of physics and mechanics to explain river processes and have tended to view statistics as secondary or complementary tools to address uncertainties in measurements and variability in phenomena such as temporal patterns (e.g. event frequency) or in attributes of physical laws (e.g. velocity, grain size). Statistical models are used in a physical approach to simplify a part of the real world, producing coefficients of physical laws without understanding or calculating all driving factors and sources of heterogeneity. Roughness coefficients produced from different approaches are a good illustration of this. This 'mix of physical arguments and pure empiricism' (Rhoads 1992) confronts geomorphologists with the problem of relating theoretical or experimental hypotheses that are usually expressed in dimensionless forms with results, and frequently introduces scaling problems. Although laboratory or field experiments help in understanding the physical laws controlling channel forms and processes, they are often uniscalar and atemporal, one would say reductionist. They are indeed unable to account for the complexity of geomorphological phenomena (controlled by climatic, geological and topographic contexts existing at the Earth's surface) along with human impacts, heterogeneous in space and time. Exploratory approaches are then very complementary to experimental approaches and a similar problem in fluvial geomorphology can be approached by upscaling and downscaling perspectives. From the grain to landscape features or from the landscape features to the grain are then two complementary strategies for understanding the relationships between forms and processes.

Examination of the literature in this field over the last two decades (1708 papers published in *Geomorphology*, *Earth Surface Processes and Landforms*, *Catena* and *Zeitschrift für Geomorphologie*) suggests that similarly to other scientists, geomorphologists are increasingly using statistics (Fig. 21.1) (Piégay *et al.* 2015). A substantial change occurred around 2002. Whereas only 15% of papers in fluvial geomorphology used statistics between 1987 and 2001, the proportion increased to 30% after 2002. In absolute numbers, reflecting the number of published papers overall, most of the contributions came from the United States and the United Kingdom, but authors in Spain, France, Japan and Belgium are the ones who most frequently used statistics (32–40% of the published papers from those countries). Although fluvial geomorphology has lagged behind its sister disciplines during recent decades in the field of statistics, this new period seems to modify the scientific landscape. This new interest in statistics takes its root within the tradition of the quantitative geomorphology of the 1960s, focusing on the linkages between forms and processes and exploring space–time framework complexity and fluvial system interactions with bivariate statistics. When searching publications focusing on 'statistics' and 'geomorphology', the Google search engine first identified Strahler (1954) and this chapter in the first edition of this book, suggesting that in recent decades, statistics were not commonly employed in fluvial geomorphic studies.

The classic text by Leopold *et al.* (1964) illustrates the use of statistics in this field in the 1950s and 1960s; 83% of graphic illustrations were bivariate scatters. The leading concepts such as drainage organization and magnitude and frequency of flow and sediment transport were based on the work of pioneering researchers, many of whom were engaged in engineering and earth sciences (Horton 1945; Strahler 1952, 1954; Wolman and Miller 1960). Statistical analyses were applied to detect correlations between variables related to each other among climate, flow characteristics and channel form and to evaluate regional controls and scale effects.

The discipline of fluvial geomorphology deals with a wide range of scales and large variability, much as faced by social sciences and biology. Heterogeneity within and between spatial units is so high that it must be assessed and understood before making progress in the field and characterizing physical processes.

Over recent decades, significant changes occurred with the emergence of new technologies (e.g. ADCP, LISST, LiDAR, ground-penetrating radar, airborne/satellite imagery, ground sensors in a wide sense, and so on), providing large datasets favouring data mining and analysis. The increased calculation and storage capacities also provided opportunities to merge

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

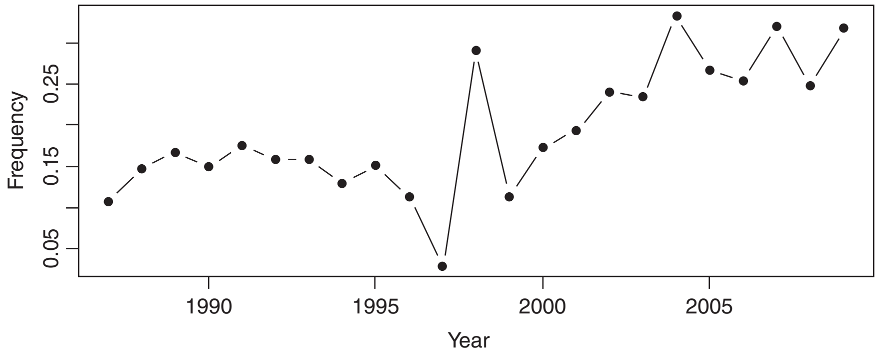


Figure 21.1 Occurrence of statistical approaches in the MS published in fluvial geomorphology between 1987 and 2009.

or pool data and explore large datasets with a multi-scale perspective. New technologies and the new information that they provide open up challenging issues for re-exploring the space–time framework at different scales from grains to landscape features and from less than one second to decades or longer. In the past, to measure hydraulic parameters intensively, geomorphologists could focus only on the local scale, whereas it is now possible to explore the reach and even regional scales (Ferguson 2008), re-enriching the hydraulic geometry approaches from the 1960s. In a sense, geomorphic questions are not only mechanistic but also focused on stochastic processes and variability. The emerging concept of riverscape is renewing spatial questioning related to issues of interest for practitioners and ecologists. Increased linkages between geomorphology and ecology, where variability is a central question, are undoubtedly promoting such approaches.

Hence the aim of this chapter is to provide a partial review of the statistical tools available, to give some examples that illustrate their use to answer geomorphological questions and to provide a simple overview of their advantages and limits. We consider fluvial geomorphology in its widest sense and incorporate consideration of the floodplain and watershed systems. We have chosen a simple organization of the chapter for illustrating the potential of different tools. Nevertheless, from a set of examples used at different times in the chapter, we shall see that it is difficult to separate the different procedures. A statistical approach, notably when exploratory, is usually conducted step by step following a set of procedures for describing variables, exploring relationships for simplifying datasets, identifying groups, modelling relationships sometimes by group, and validating and assessing accuracy, precision, bias or errors.

Interest of statistical tools for fluvial geomorphologists

Statistics can be defined as a set of mathematical techniques used to collect, characterize, summarize and classify numerical data, identify groups or test differences between them, detect correlations between variables and provide predictions. They also assess the errors, uncertainties and accuracy that come with these results. They are commonly used to interpret phenomena for which an exhaustive study of all the acting factors and populations is not possible owing to their great number or their complexity. Statistical analyses allow the characterization of large populations through the information collected on a limited number of individuals (or samples). Fluvial geomorphologists deal with complex spatial components, such as in-channel features, channel beds and reaches, valleys, watersheds, regions and even continents, whose characteristics, occurrence and spatial distribution change through time. They are also concerned with processes, mainly bedload transport, suspended sediment concentrations, flow hydraulics or vegetation dynamics, which are also variable in time. Each of them can be characterized by attributes, called 'variables', whose values can be numeric (magnitude or rank; ratios; intervals) or nominal (qualitative).

Individuals from a population of interest can be described and compared based on one or several variables. It is then possible to group them, to identify trends or patterns in space (e.g. distance downstream) or in time (e.g. measures taken at a given time interval). The systems under study might actually be viewed as spatio-temporal frameworks. The first need is often to evaluate the basic features of the variables and to summarize them. Assessing data distribution through a few summary measures (or statistics), assessing, in particular, the central tendency from the mean or the median, the variability from quantiles or standard deviation and the shape of their distribution (e.g. asymmetry), is the aim of descriptive statistics. The examination of the variables' distribution is often a preliminary step to further study as many standard statistical procedures assume that data follow the normal law of distribution – sometimes called a bell-shaped curve – or that the samples are large enough to make this assumption optional. Data distributions usually can be compared with known distributions through graph visualization or various dispersion parameters (Fig. 21.2).

A serious limitation of descriptive statistics is that they only apply to the data collected and cannot be inferred to the populations under study. Inference refers to the generalization of characteristics of samples to the total population. It is based on tests and models that validate or invalidate an *a priori*

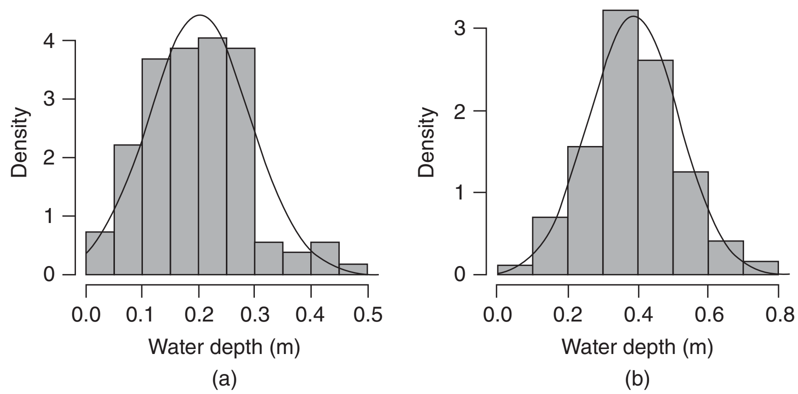


Figure 21.2 Distribution of water depths measured in the braided channels of Bès River and associated normal distribution fits for 2008 (a) and 2010 (b) hydrological conditions. Source: Tacon *et al*, 2014. Reproduced with permission of Elsevier.

hypothesis regarding the phenomenon under study (the null hypothesis H_0). Further, modelling allows us to explain and even predict environmental phenomena or features. Hence the use of descriptive statistics, especially univariate statistics, is generally just a first step towards the use of more thorough tools, e.g. inferential statistics or modelling.

It is not our objective to give a detailed summary of the statistical possibilities, as many statistical textbooks and software guidelines or websites offer a wide range of statistical options, potentially useful for environmental scientists. The main issue is to show how statistical tools can help assist in solving a geomorphological problem, in formulating and validating hypotheses and in highlighting advantages and disadvantages of various approaches.

21.2 Bivariate statistics to explore patterns of forms and their drivers

Bivariate statistics, and regressions in particular, have been one of the most popular statistical tools in geomorphology, notably in the 1960s when quantitative geomorphology emerged. They focus on the relationship, or correlation, between two variables.

Studying a numerical variable according to another one: regression analysis

Simple regression analysis has commonly been used to analyse channel form, especially since the influential work of Leopold and Maddock (1953) and is still widely employed over the last three decades of the 20th century [as illustrated by the textbooks of Gregory and Walling (1973) and Bravard and Petit (1997)].

Regression models are generally linear and often imply prior transformation (for instance, log-transformation) of the variables (see Tables 21.1 here and 20.5 and 20.6 in corresponding chapter in the first edition of this book). Other types of regressions have rarely been used. Such is the case of polynomial regression, probably because of the limited confidence in the corresponding coefficients, although graphical interpretation might help. Regression has some advantages over other methods such as pedagogical efficiency (the scatter plot gives an instantaneous view of the results), simplicity of the technique and possible use in both a predictive and explanatory way. Residuals or dummy variables are also considered complementarily to explore threshold conditions and discriminate groups. Regression strength is usually measured by a correlation coefficient. When multiple variables are compared with each other, some authors may then provide a correlation matrix rather than showing each of the regressions, as illustrated by some recent references (Table 21.1).

Among the most popular explanatory approaches explored by geomorphologists is the hydraulic geometry and the associated theory of dominant/bankfull discharge as critical drivers of geomorphic features. Channel geometry has often been described by simple regressions, such as the classical set of relationships of Leopold and Maddock (1953). These relationships between discharge and width or depth or velocity were reconsidered exhaustively by Rhoads (1992) in the light of different bivariate models and submodels. Validity conditions and criticism of the various estimation procedures were linked to measurement constraints. Such a classical approach has been applied more widely to explain geomorphic variables such as knickpoint migration rate, plunge pool depth, number of pools, meander (e.g. length, wavelength, amplitude) or valley geometry, usually relating them to basin area or discharge, to explore spatial organization and critical processes.

Regressions may be used as explanatory models, such as bivariate models of sediment concentrations according to hydrograph (Kunhle 1992) or flood sequences (Park 1992). For instance, the regressions between sediment discharge (bedload, sand, mean travel distance) and some flow characteristics can be used to define minimum flow conditions for bed material motion. Hydraulic variables may be discharge, power or bed shear stress (Shields 1936; Reid and Frostick 1986; Gomez and Church 1989; Kunhle 1992; Wilcock 1993). Assessing motion thresholds by particle diameter classes is rarely successful because particles do not move solely according to their size but also according to other criteria such as size mixture, shape

Table 21.1 Describe and assess the link between variables through regressions and the differences between groups through tests – some examples.

Aim	Example of statistical tools	Examples of application in fluvial geomorphology	References
Describe and test the link between two variables through regressions	Pearson's correlation coefficient	Channel adjustment versus aquatic habitat characteristics	Mazeika *et al.* (2004)
	Spearman's correlation coefficient	Parameters describing braiding versus control factors (exceedance flow frequency and normalized active channel width)	Belletti *et al.* (2012)
	Simple regression	See Tables 20.5 and 20.6 in the first edition Width versus discharge, depth versus discharge, stream power versus discharge (power function)	Fonstad and Marcus (2010)
	Multiple regression	See Tables 20.5 and 20.6 in the first edition Channel bankfull dimension and shape, hydraulics, bedform wavelength and amplitude, grain size, flow resistance, standard deviation of hydraulic radius, and volume of large woody debris, versus potential control variables (drainage area, discharge, bed gradient). (Power and linear regression)	Wohl *et al.* (2004)
		Effects of tributary flux ratio (*FR*), flux calibre ratio (*DR*), and discharge ratio (*QR*) on the response of a 10 km concave mainstream (in terms of slope, grain size, elevation)	Ferguson *et al.* (2006)
		Grain size prediction from aerial images	Verdu *et al.* (2005)
	Polynomial regression	Sample density versus depth of sampling	Reneau and Dietrich (1991)
		Proportion of channel width before and after cut-off versus diversion angle	Constantine *et al.* (2010)
Describe and test differences between groups in variables through parametric tests and models	Student's *t*-test	Channel width and depth at two dates	Rhoads and Miller (1991)
		Median grain size measured by three operators	Wohl *et al.* (1996)
		Grain size measured at different sites	Dawson (1988)
	Linear model (ANOVA)	D_{50} and D_{84} measured by three operators	Wohl *et al.* (1996)
	Post-hoc analysis	Differences in the species richness and number of viable propagules between reaches and sampling periods	Gurnell *et al.* (2007)
Describe and test differences between groups in variables through non-parametric tests and models	Wilcoxon signed-rank test	Residuals of the regression "Q2 versus catchment size" and a set of other hydromorphic indicators compared to 2 classes of reach (urban versus reference)	Navratil *et al.* (2013)
	Kruskal–Wallis test	Channel vertical changes versus number of mining sites, number of upland active torrents, ratio of eroding banks	Liébault *et al.* (2013)
	Chi-squared test	Grain size distributions (classes)	Wohl *et al.* (1996)
	Kolmogorov–Smirnov	Distributions of source and tributary source link lengths	Knighton *et al.* (1992)

and bed structure or position within the longitudinal profile. This masking influence of unaccounted for factors is also responsible for the scatter in relations between particle size and travel length.

Regression techniques are also commonly used to define predictive rating curves relating a quantity that requires a long time and field effort to be quantified (e.g. suspended sediment concentration or bedload transport rate) to a variable that is relatively easy to record (e.g. discharge or stage). Such statistical models have thus been applied to predict geomorphic processes such as floodplain sedimentation rates, sediment delivery ratio, channel shifting related to catchment size or other geometric factors.

Residuals

Although coefficients can be statistically valid because of large samples, relations are often blurred by data scattering, which poses the problem of residual interpretation and of further processing. Sometimes larger residuals are commented upon and eventually withdrawn from regressions if found to be exceptions (Lecce 1997). The discrepancy between observed and fitted values (i.e. residuals) may yield important insights, for instance, if different trends appear on two sides of a threshold. High positive residuals may thus indicate some radical change in flow dynamics such as the destruction of the armored bed layer above a discharge or power value close to or at bankfull stage (Batalla and Sala 1995; Batalla 1997). Large, and even anomalous, residuals may exhibit some time trend or hysteresis effect. This is common in sediment rating curves and reflects temporal variations in sediment availability during flood events or on a seasonal basis.

Grouping

Datasets can be partitioned according to various criteria, such as distance from sources**,** bank deposits **or** vegetation (Hey and Thorne 1983; Ferguson 1986a). Regression models might show differences according to some grouping of the individuals linked to environmental conditions, such as bank material, vegetation types, planform (e.g. braiding and meandering) or lithology in the case of the power function relating width–depth and discharge (Schumm 1960; Ferguson 1986a). To assess these effects, individual data points can be represented differently on the regression plot (or on the graph of residual values) according to the group to which they belong. For example, the power equations linking width with discharge have different coefficients for riffles versus pools, while the exponents are very similar, requiring distinct models for the two groups (Richards 1976). Petit (1987) and Sear (1996) have also related discharge to shear stress and bedload transport to stream power, respectively, according to whether points are located within pools or riffle sections: where the regression lines cross defines the threshold at which transport becomes more efficient in pools. Sear (1996) also illustrated this efficiency by the relation between mean distance travelled by particles >20 mm and excess stream power and used the Shield's entrainment function to demonstrate higher entrainment thresholds over riffles (Fig. 21.3). Dispersion of data as shown by residuals from the regression line was explained by the effect of different textural and structural features of the bed sediments, which changed as discharge increased and flow type changed, and sedimentological differences such as grain size, bed strength, structure and cluster components were demonstrated through Mann–Whitney tests for population differences. Distinct populations can be distinguished by discriminant lines, such as the slope–discharge plot on which braided versus meandering river patterns were distinguished by Leopold and Wolman (1957). These categories are considered to be discriminant variables although corresponding statistics are not employed. Selection of bounds is consequently subjective while degrees of freedom are reduced. In a more general way, residuals can be classified through cluster analysis.

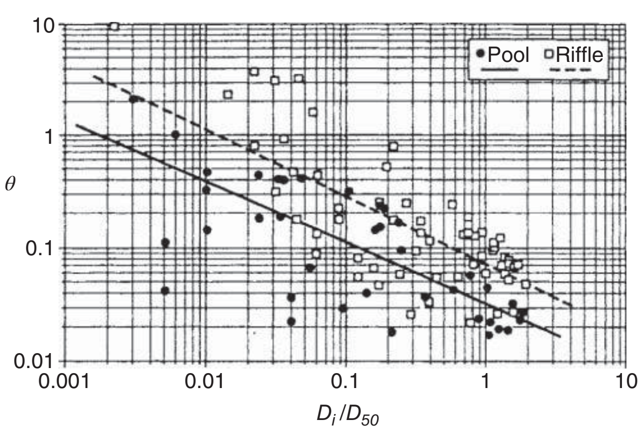


Figure 21.3 Bivariate plot linking the Shields parameter θ with the D_i/D_{50} ratio. Two models were fitted, one for the pools and one for the riffles observed in the North Tyne. Source: Sear, 1996. Reproduced with permission of Wiley.

Critical appraisal

Regression plots can be very illustrative and allow exploration of causalities in term of interactions between response and controlling variables, but they are easily misused or misinterpreted in different contexts. Models defined and fitted in a given local or regional context cannot be applied to other regional contexts outside their validity domains. For instance, Brookes (1987) examined the relation between stream power and channel sensitivity for a set of stream projects and estimated that no channel change occurred at stream powers less than 35 W m^{-2} for the studied setting, but this would not be justified as a simple rule of thumb in other regional contexts, because of differences in channel resistance, which affect the threshold. For example, sand-bed rivers can adjust to unit stream powers less than 35 W m^{-2}, whereas some river geometries cannot adjust even to stream powers far above 35 W m^{-2}.

Another criticism of regression misuse is due to the frequent confusion between correlation (or co-occurrence) and causality. Linear regression is a common example in which assumptions are made about the scatter distribution of the probable value of the dependent variable corresponding to a single value of the independent one. Distinction between dependent and independent variables may be hypothetical and is often based upon qualitative field knowledge. The interaction of fluvial variables makes the latter distinction somehow artificial so that other names are preferred, such as *response variable* and *regressor*. Indeed, slope, width, discharge and width are intercorrelated, hence approaching causality with such variables can cause problems. For example, in the Leopold and Wolman (1957) plot, braided rivers exhibit a higher slope than meandering rivers. However, this pattern is actually due mostly to substantial bedload delivery occurring in uplands, where slopes are high. Braided patterns are probably not caused by high slopes, although they occur in contexts where slopes are high. In fact, they occur where there is a significant decrease in slope (from that of the channel upstream), limiting the transport capacity of the river and inducing deposits of sediment provided by its upstream steep tributaries.

Some authors have included normalized or dimensionless data to produce better fits or better agreements with theoretical statements (Church and Mark 1980). Most of the relations follow a log–log or semi-logarithmic form (Table 21.1). Consequently, correction factors should be applied as when using detransformed log functions (Ferguson 1986b). However statistical bias may be introduced through transformations, especially when data are not homogeneous in their distribution. The use of dimensionless variables may cause spurious correlations if some common scaling appears on both sides of the equation (Rhoads 1992).

Specific case of multiple regression

Approaches based on simple regressions are sometimes too reductive to assess geomorphic causalities, so that multiple regressions have also been used to highlight geomorphic relationships or establish predictive models. Such regressions can explain and predict the values of a quantitative variable according to several explanatory variables if sample sizes are large enough. Many possibilities exist to build such models, such as the standard linear multiple regression ($Y = a_i X_i + b$), possibly implying a prior transformation (for instance, log transformation) of the variables. Multiple regressions sometimes follow a stepwise procedure, which helps select the relevant explanatory variables and establish a hierarchy among them. Standardized regression coefficients β are used to order variables according to their respective contributions. Categorical variables can be introduced in order to improve multiple correlations. They take the value 1 (present) or 0 (absent) in the regression. In most of the geomorphic contributions, categorical variables are usually specific sub-units such as the location of cross-sections with respect to checkdams (upstream/downstream) to assess their geomorphic and phyto-ecological effects, for which Bombino *et al.* (2009) established multiple regression relationships between components summarizing geomorphic characters (e.g. slope, median grain size, channel shape, sub-surface fine sediment content) provided by a preliminary principal component analysis (PCA) and including locations and vegetation parameters (vegetation extent, development, cross-sectional variability, number of species present). Hence multiple regressions allow us to gradually complexify simple regressions to take into account extra factors. For instance, along with his study of bivariate models linking discharge and width or depth or velocity, Rhoads (1992) also compared multiple regression models of channel geometry adjustments and introduced variations of discharge and bed material properties as regressors. Walling and He (1998) established an exponential model predicting the floodplain sedimentation rate from the lateral distance to the channel according to the flood depth and the mean sediment concentration (Fig. 21.4). Rickenmann (1997) used multiple predictive equations to relate total bedload transport to water volumes and peak discharges over a threshold of $0.5\,m^3\,s^{-1}$ in pre-Alpine Swiss watersheds.

Identification, selection and transformation of variables are often difficult although essential steps in carrying out a multiple regression. As an example, valley widths appeared to be more influential than mean stream power in the statistical explanation of post-European settlement alluvium in the Wisconsin Driftless Area (Lecce 1997). However, more insights were possible through a combination of other qualitative and quantitative relationships linked to the existence and functioning of meander belts in medium-sized tributaries. To cite another example, relationships established between the peak flow and the suspended sediment concentration can be improved by adding supplementary variables such as the season, or the falling or rising flood stage, to produce more powerful models. However determination coefficients in multivariate regressions may not increase significantly because of wide scatter in the data or the introduction of inappropriate variables. Some important influences are difficult or impossible to measure in the field, e.g. bed structure (Hassan *et al.* 1992), velocity pulses (Hoey 1992) or roughness of migrating bed forms during floods.

Describing and testing differences in a variable between groups

One of the main tasks in fluvial geomorphology is to distinguish spatial entities (bars, channel reaches, floodplain features, sedimentary facies) according to their specific characteristics, identifying critical thresholds or drivers explaining differences between spatial entities. In some cases, the groups are defined prior to the analysis and one wishes to describe the differences between them. In other cases, the groups are not predefined, but are identified as a result of the analysis. In the former cases (groups are defined prior to analysis), one might wish to describe the difference between groups and test that the difference observed on data derives from an actual difference between the studied populations. This can be achieved through the use of many kinds of tests, either parametric or non-parametric. The term ‘parametric’ applies to methods that assume either that the variables follow a standard distribution or that there are enough data to consider that the results of the parametric method are approximately accurate even though the distribution is not standard. In contrast, ‘non-parametric’ refers to methods that make no assumptions as for the distribution of variables.

Parametric or non-parametric tests can be selected according to the type and the size of the sets, but also the shape of the distribution of the variable (normal, log-normal, etc.). For numerical, continuous variables, parametric tests such as the *t*-test (Student) or analysis of variance (ANOVA) might be used to compare the mean value of a variable according to the modality of a categorical variable (defining groups). For example, an ANOVA distinguishes two parts: the variability of the measure within each group (e.g. a set of reaches) and the variability of the measure between the groups (e.g. between different meaningful sets of reaches). The greater this second variability is relative to the first, the more significant is the difference between the groups.

In the case of nominal and ordinal variables, non-parametric tests on a pairwise basis might be used to compare the distributions of groups (Table 21.1). Although they have no underlying distribution assumptions and do not require large samples, these tools are rarely encountered in the geomorphological literature, probably because of their lower concluding flexibility and power. The topological characteristics of networks are one field of application for non-parametric tests. To assess the role of local disturbances on bed sediment distribution in low-order streams, Rice and Church (1996b) tested the hypothesis that the systematic downstream reduction of grain size – the negative exponential model – is precluded by colluvial inputs and log jams whose distribution is random in both space and age. An ANOVA showed significant differences in surficial D_{50} among study reaches and established the textural differences between a reach decoupled from lateral slope inputs and one that was not.

The Kruskal–Wallis non-parametric test (which is the equivalent of ANOVA used for parametric approach) was used to test

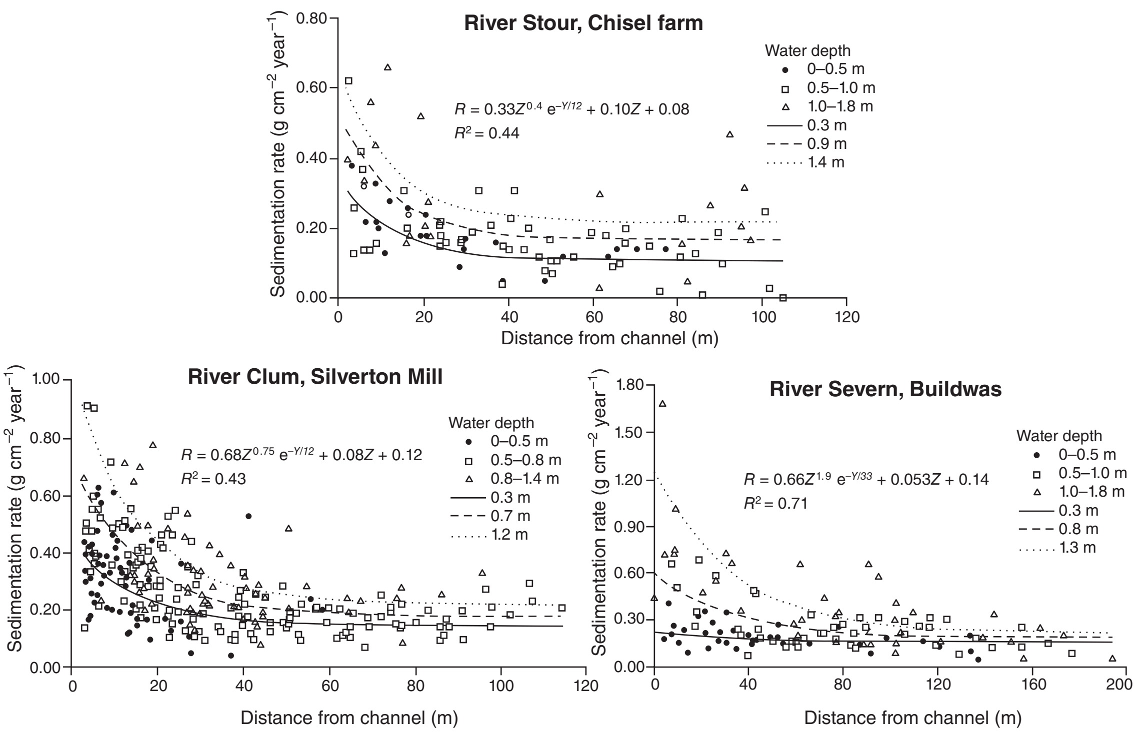


Figure 21.4 Prediction of floodplain sedimentation rate from the lateral distance to the channel according to flood depth and mean sediment concentration. Source: Walling and He, 1998. Reproduced with permission of Elsevier.

potential factors controlling braided channel vertical adjustment related to pluri-decadal changes in sediment delivery and demonstrated that long-term altimetric changes were significantly related to the number of active tributaries, erodability of banks and the number of gravel-mining sites (Fig. 21.5) (Liébault *et al.* 2013).

The Kolmogorov–Smirnov (KS) test is also used in some other contexts, being sensitive not only to the difference of means but also to differences in the shape of the distribution of two compared variables. Knighton *et al.* (1992) compared the size and extent of tidal channels of northern Australia in different years using the KS test, thereby documenting that network evolution through time followed an exponential growth.

Such an approach is also used when considering two nominal variables for testing their independence based on a contingency table. We used chi-squared tests to assess the difference in grain size distribution between sites on two streams of the Massif Central (France) based on grain size visual classes. We compared the morphologies and grain sizes of three reaches to evaluate the potential effects of in-channel wood storage on sediment deposition. Site D_1 had no wood, whereas D_2 averaged 20 kg per metre of river length and D_3, located immediately downstream of D_2, averaged 38 kg m^{-1}. We randomly sampled the bed and determined the dominant grain size in each reach. The hypothesis was that grain size distribution was different between reaches as a function of in channel wood abundance. The chi-squared test confirmed that the three grain size distributions were different, with D_1 the most heterogeneous and D_2 and D_3 having a higher frequency of one or two classes. D_2 had a high frequency of sandy plots associated with side channel jams, whereas D_3, within which wood formed jams, had bars composed of 8–32 mm gravel (Clément and Piégay 2003).

When considering field geomorphology, the links between factors are often complex and noisy because other variables than those being tested are acting. For instance, unexpected human pressures and associated river adjustment might occur or the environmental setting might be more complex than anticipated in the sampling design. Moreover, some of the variables require a significant field effort (e.g. bedload transport value) so that the number of observations is often low. In such conditions, it is often difficult to apply classical statistical tests that have rigorous requirements, such as minimum number of individuals, equality of variance or normal distribution of variables. Resorting to unparametric tests to minimize such requirements is then appropriate.

21.3 Exploration of datasets using multivariate statistics

Describe a dataset

In case one is faced with a whole dataset and has no *a priori* insights as to which are the key variables to understand the

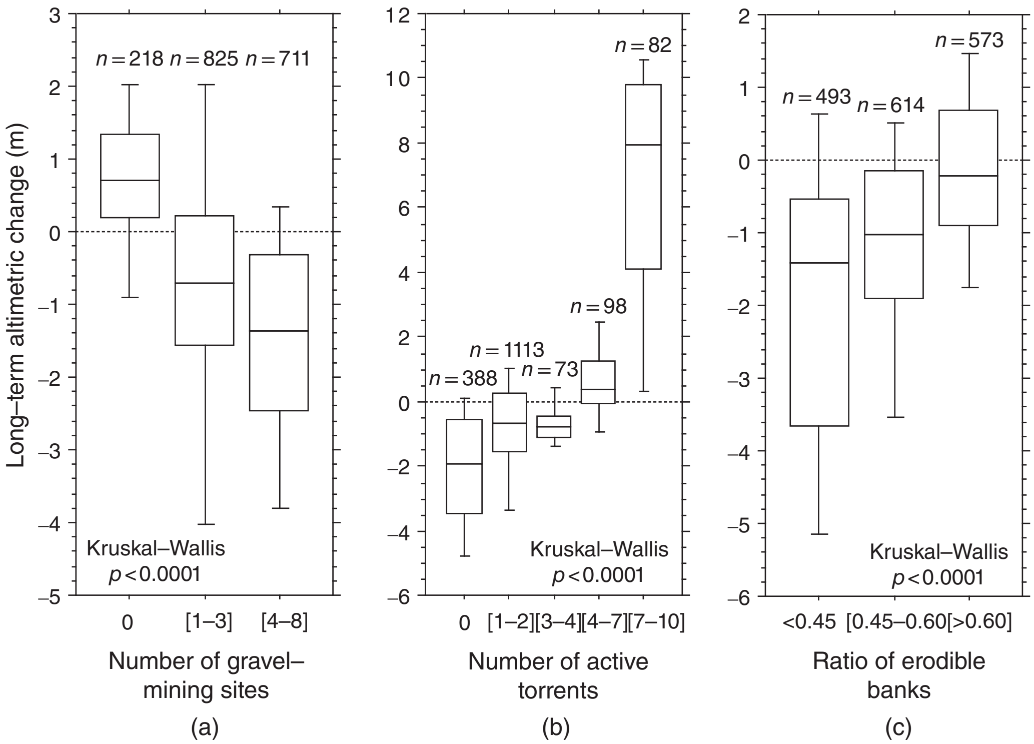


Figure 21.5 Box plots of the distribution of long-term altimetric changes as a function of (a) the number of gravel-mining sites, (b) the number of active torrents and (c) the ratio of erodible banks; *n* refers to the number of sub-reaches in each group; boxes represent first and third quartiles, horizontal lines in the boxes represent the medians, vertical lines above and below the box the first and ninth deciles. Source: Liébault *et al*, 2013. Reproduced with permission of Wiley.

phenomena at stake, factorial analyses can be good exploratory tools. Indeed, with recent increases in computational capabilities, they can now summarize the structure of large datasets. Factorial methods compute the main axes of multidimensional scatter plots and produce simplified graphics of the descriptive elements identified. Depending on the data characteristics (continuous versus categorical variables), various techniques are used including Principal Component Analysis (PCA), Correspondence Analysis (CA) and Multiple Correspondence Analysis (MCA) (Table 21.2).

Among factorial techniques, PCA is the most common. It examines a set of continuous attribute variables (e.g. channel width, depth, grain size, sinuosity) measured at different sites (e.g. cross-sections, reaches) and identifies the key associations between them by reducing a large number of correlated variables to a smaller, more manageable set of factors. The principal components can be defined as new variables that summarize the information carried by the set of correlated measured variables. CA is a weighted PCA of a contingency table, whereas MCA examines the relations between categorical variables that are reduced to dichotomous variables (absence versus presence).

For instance, multivariate analyses can simplify datasets such as morphometric properties of gullies in Guyana savannas (Ebisemiju and Ekiti 1989) or geochemical signatures of surficial deposits in northern England (Passmore and Macklin 1994). In the last example, because elemental compositions, notably heavy mineral concentrations (Pb, Zn), depend on geological conditions and historical mining operations, this descriptive approach allows deposits to be distinguished according to their provenance from geologically distinct sub-catchments. CA was also used by Riquier *et al.* (2015) to show how different restored floodplain channels are, in terms of sediment conditions (e.g. grain size classes, overbank sediment covering, coarsening versus fining along restored channel). Belletti *et al.* (2013) compared 50 braided channels across the French Alps to understand how their features adjust to hydro-climatic contexts.

Moreover, multivariate analyses can also be carried out to describe what makes groups or individuals different from each other. Such an exploratory approach might consist in identifying the common structure in two sets of variables – *a* and *b* – characterizing a similar set of stations – *S* – through a CoInertia Analysis or a Canonical Correspondence Analysis. Another way to describe differences between groups is to provide principal components that discriminate groups as clearly as possible. This can be achieved through discriminant analysis.

Explore the co-structure of two datasets

The Co-Inertia Analysis searches the Co-Inertia axes that maximize the covariance of projection coordinates of the two datatables for which each structure has previously been studied

Table 21.2 Explore datasets through factorial analysis to highlight relationships between variables or group variables or individuals – some examples.

Aim	Example of statistical tools	Examples of application in fluvial geomorphology	References
Describe a dataset	PCA, CA, MCA	Magnetic properties of sediments	Yu and Oldfield (1993)
		Mineralogy characteristics of sediments	Llorens *et al.* (1997)
		Geochemical concentrations in sediment samples	Passmore and Macklin (1994)
		Aquatic channel pattern within braided reaches	Belletti *et al.* (2012)
		Sediment pattern in restored former channels	Riquier *et al.* (2015)
	Multi-dimensional scaling	Textural and geochemical sediment characters at different spatial scale	Thoms *et al.* (2007)
Highlight co-structure between two datasets	Canonical correspondence analysis,	Drainage basin characteristics versus potential environmental controlled factors	Ebisemiju (1988)
	Co-inertia analysis	Field characteristics of former channel plugs versus former channel/channel characters	Piégay *et al.* (2002)
		Channel morphology versus basin characteristics	Liébault *et al.* (2002)
Distinguish groups of individuals	Discriminant analysis	Set of channel indicators versus disturbed/undisturbed reaches	Woodsmith and Buffington (1996)
		Morphological descriptors versus reaches	Gurnell (1997)
		Morphometric variables of meander pattern versus meander types/models	Howard and Hemberger (1991)
		Hydraulic geometry descriptors versus reaches according to bank stability	Ridenour and Giardino (1995)
		Radiometric and geometric imagery index versus different in channel features pools, riffles)	Wiederkehr *et al.* (2010)
Distinguish groups of individuals through clustering methods	Fuzzy clustering *k*-means clustering	Sediment source tracing based on soil magnetic data	Hatfield and Maher (2009)
	Hierarchical clustering	Dissimilarities between physical habitat characteristics	Gurnell *et al.* (2007)
		Braided river types according to aquatic channel patterns	Belletti *et al.* (2012)
		Restored former channel types according to sedimentation characters	Riquier *et al.* (2015)
		Typology of braided rivers	Piégay *et al.* (2009)

with factorial analysis. It can be used, for example, to identify for given river reaches a co-structure between two groups of variables, one describing channel and the other floodplain (for details, see Tucker 1958 or Chessel and Mercier 1993). A Co-Inertia Analysis was carried out along the Ain River, France (3640 km^2) on 'terrestrial plugs', floodplain areas separating the main channel from the permanent aquatic zone of former channels such as oxbow lakes (Piégay *et al.* 2002). We built two datasets: (i) field data describing the floodplain biogeomorphology and (ii) large-scale structural data based on aerial photography analysis and historical documents describing the environmental changes (e.g. main channel aggradation, degradation, shifting) and dating fluvial forms (e.g. cut-offs). This analysis identified the co-structure of environmental variables and field variables. Figure 21.6a shows both the 'plug' field plots (end of the arrows) and former channels (circles) on the first factorial map. Figure 21.6b displays the modalities of the structural and field variables on the first Co-Inertia factorial map. Two main groups of plugs can then be distinguished according to their original geomorphic pattern (STY): the plugs of the former meandering channels and the plugs of the former braided channels. Other variables, such as the angle between the main channel and the former channel (ANG) or channel shifting (HOI) and degradation (VEI), do not exhibit clear patterns on the first Co-Inertia factorial map. The difference between the two groups defined above is also related to age (i.e. gradient from old to medium-aged). Among the braided former channels, the oldest ones exhibit fine grain size, thick overbank deposits and numerous sediment facies. Their vegetation includes both hardwood and semi-aquatic communities. The youngest (PN1/2, CFO) have characteristics similar to those of meanders: fine to medium grain size and moderately thick overbank deposits.

In the Canonical Correspondence Analysis, the link between two sets of continuous variables is tested. This technique has similarities with Co-Inertia Analysis described previously, but is less flexible and not as easy to interpret because inferential assumptions are required. If variables can be divided into two sets, canonical correlation provides a suitable simplifying tool as the model successively finds pairs of linear combinations from each set (canonical variables) such that the correlation between the canonical variables is maximized. Each of the canonical variables is uncorrelated with all the similar variables in the other pairs. The analysis can go on so as to find other sets of canonical variables uncorrelated with the first pair, the limit of combination being the number of variables in the smaller set. In an example from eastern Nigeria, morphometric properties of subcatchments were related to their relief, soils and vegetation cover (Ebisemiju 1988). Canonical Correspondence Analysis was used to distinguish three patterns of association: texture of dissection versus soil and vegetation characteristics, network size versus stage of basin relief evolution and bifurcation ratio versus basin relief. Simplification and independence of identified patterns are certainly advantages but, as in other methods, fulfilling the statistical requirements of the method, such as normality and multicolinearity of the data, may be problematic. As in other multivariate analyses, interpretation can be difficult owing to the intricate relations between variables, which are sometimes ambiguous, especially if they are indices or ratios.

Identify groups within a dataset

Discriminant analysis produces a factorial map maximizing the ratio of between-group variance to total variance. Such analyses might be more thorough than purely descriptive analyses; in particular, discriminant functions can be established and used for predictive purposes when using additional data. To identify geomorphic factors (slope, grain size, channel width) controlling in-channel features (pool–riffle, step–pool and plane–bed) in a worldwide dataset of mountain streams, Wohl and Merritt (2005) used discriminant analysis. Similarly, Owens *et al.* (1999) used multivariate discriminant analyses to identify the optimal combination of geochemical and mineral magnetic properties for discriminating contributing sources

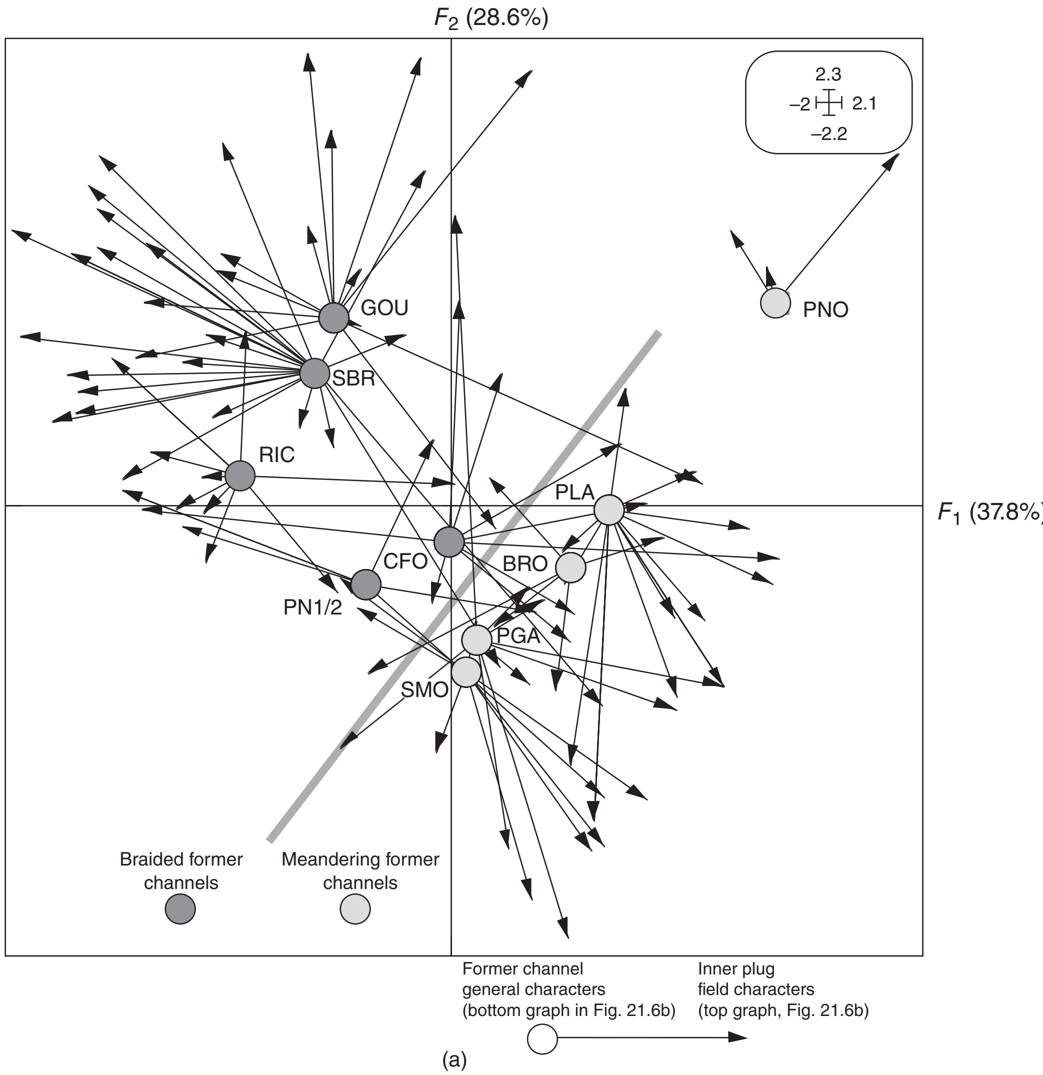


Figure 21.6 Co-Inertia Analysis between field data and environmental data describing the plugs of the former channels of the Ain River. (a) Match of the two scatters of the first factorial map; (b) projection of the modalities of the environmental and field variables on the first factorial map. Source: Piégay *et al*, 2002. Reproduced with permission of Wiley.

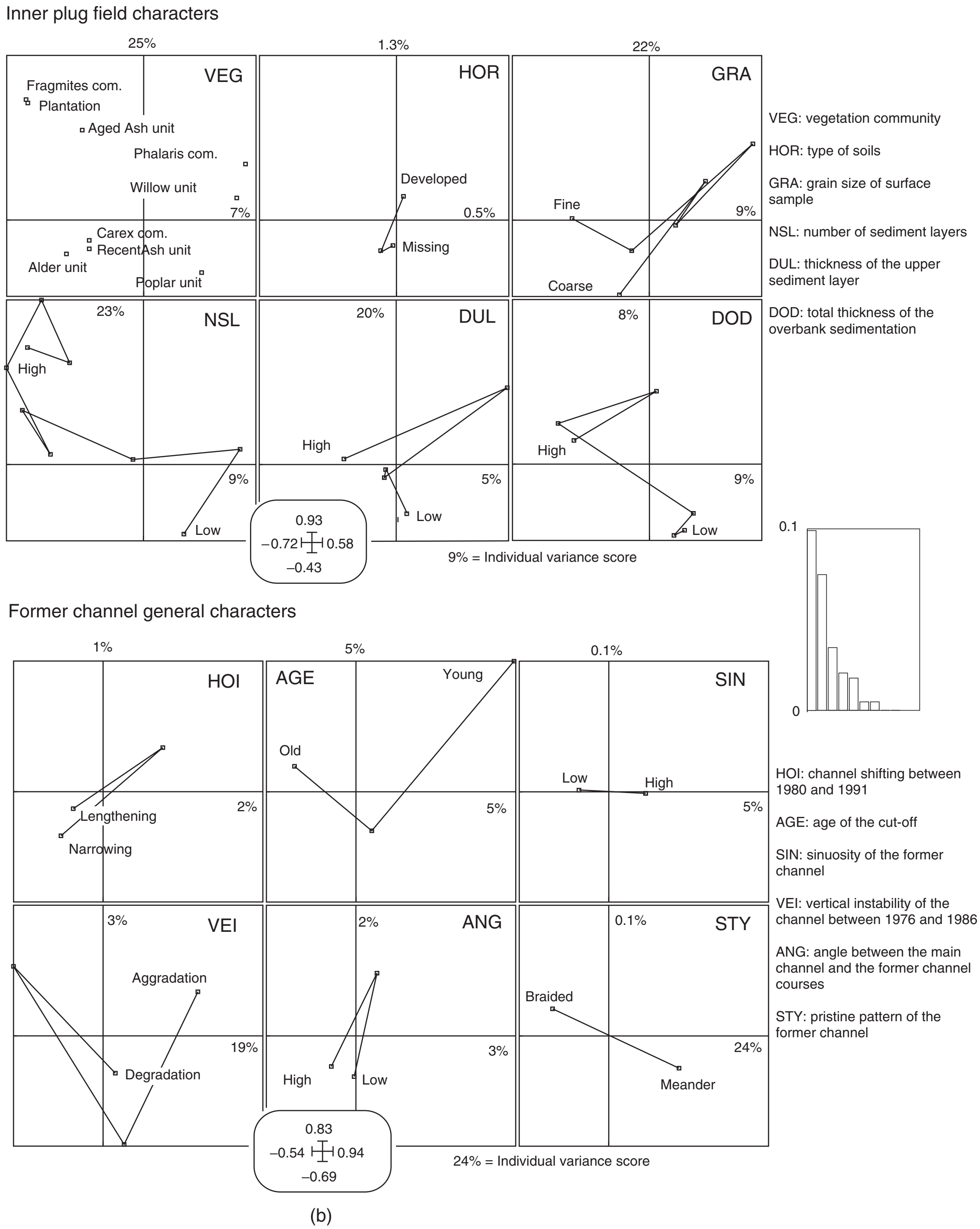


Figure 21.6 (*continued*)

to floodplain sedimentation (topsoil versus subsoil/channel bank; main geological zones) in the Ouse River basin in England. Wiederkehr *et al.* (2010) used such techniques to extract geomorphic features from airborne imagery and test which were the critical radiometric and geometric variables having the best discriminating capacity. Pools, riffles and lentic/lotic channels were then well differentiated through the calculation of a discriminant function, which allowed the prediction and mapping of various habitat types.

Clustering methods (e.g. *k*-means method or hierarchical classifications) assign individuals or samples to groups (or clusters) so that the objects in the same cluster are more similar to each other than to those in other clusters, using similarity or distance computation algorithms to distinguish groups. They are not based on any distributional assumption and may be used when the variables are not independent. Factorial and cluster analysis are often complementary, as factorial analysis can be used first to simplify a dataset and clustering methods can then be used to distinguish groups based on the principal components. For example, Emery *et al.* (2003) detected distinct hydraulic features within a channel reach based on velocity measured at three flow stages. Nelson *et al.* (2014) tested four cluster analysis methods (e.g. *k*-means algorithm, spatially constrained agglomerative clustering, spectral clustering, fuzzy clustering) to identify grain size patches in flume experiments and concluded that all methods produced better grain size patch than a visual assessment.

At a larger scale, Parsons and Thoms (2007) studied the interactions between channel morphology and in-channel wood characteristics based on a preliminary cluster analysis, which identified functional geomorphic features from meanderbelt width, meander wavelength, amplitude, length, width, tightness and angle. Similarly, dendrograms showing braided reaches were clustered according to their distances based on different descriptors, as shown in Fig. 21.7. Box-plots are then used to interpret the braided characteristics of each of the classes. Type 1 is a set of braided rivers located in lowland areas and characterized by fairly low slopes, narrow channel and with a wide forest corridor. Some interpretations can be established at a regional scale in term of critical drivers (intensity of sediment delivery, hydrological context) or adjustment state through a temporal trajectory (Piégay *et al.* 2009).

Explaining and predicting geomorphic relationships through multivariate analysis: an often composite process

To study an environmental phenomenon or process thoroughly usually involves several statistical steps. Initially, factorial analyses can be powerful tools to discriminate sampling sites, but also to define deterministic models for prediction (see fig. 20.5 in Clément and Piégay 2003). Once the large dataset has been summarized by factorial analysis, further analysis of the components can be done. Indeed, through the exploratory description of datasets that they offer, factorial analyses can be used to distinguish between two sets of individuals or spatial entities, and to identify the variables that are strongly inter-correlated, suggesting causal links, hence providing insight to define the models.

One example of such a composite analysis deals with French southern pre-Alpine mountain streams. One objective was to predict channel narrowing, observed over recent decades on these streams, based on other characteristics, such as geometry and grain size, and also to identify factors controlling such processes at the catchment scale (see Chapter 17). The model was defined using data from the Eygues basin but was used to predict channel narrowing of the tributaries of the Drôme and the Roubion, two neighbouring rivers. First, a normalized PCA was used to synthesize the channel parameters (Fig. 21.8a). The first map constituted a good summary, with the first component distinguishing wide channels with fine bed materials (e.g. Sauve, Bentrix, Rieu Sec), from narrow, coarse-grained channels, while the second component distinguished steep channels with poorly sorted beds from deep channels. The first two components were then power-transformed $[(X + 5)^{-0.5}]$ and used as regressors to predict the channel narrowing observed between 1945 and 1995 on aerial photographs. The scatter plot of observed versus predicted channel narrowing (Fig. 21.8b) showed that narrowing mainly occurred in reaches characterized by high embeddeness, coarse grain sizes and steep gradients. The geomorphic interpretation is that narrowing is observed in high-energy tributaries, with narrow valleys located closer to the basin sediment sources. These channels first experienced a decrease in bedload supply with land use changes in the basin in the early 20th century. The sediment moving into the channel was then rapidly exported, the channel was slightly degraded and the coarse bars were colonized by vegetation. Because they have a limited capacity to store gravel, they are mostly conveyor channels rather than depositional reaches. These channels are now narrowed, slightly degraded, paved and embedded. Unlike the sequence of channel incision and widening observed in the loess region of the Mississippi inner delta, channel incision in the French pre-Alps is associated with channel narrowing. Grain size and channel form parameters were also measured on the Drôme and Roubion tributaries and they were added to the normalized PCA as supplementary stations. This approach did not change the previously calculated factorial map but projected new individuals on it. We then predicted their narrowing using the model performed on the Eygues tributaries and compared these values with the observed values. The model fitted well, suggesting that it may be broadly applicable across a large geographical region with roughly similar hydrology, geomorphology and land-use history.

21.4 Describing, explaining and predicting through probabilities and distributions

When studying a random variable, the output might correspond to a metric (e.g. mean value) of a given variable, but also to a probability or a distribution (either empirical or a known parametric distributions such as normal, exponential, gamma distributions, etc.). Probabilities are also useful to generate models in which the variables of interest are categorical, such as indicator variables of events (e.g. occurrence of peak flows).

Explaining and predicting probabilities of events: logistic and multinomial models

One of the most basic tools dealing with probabilities are the logistic or multinomial models. They generalize the principles of linear models (such as those corresponding to regressions

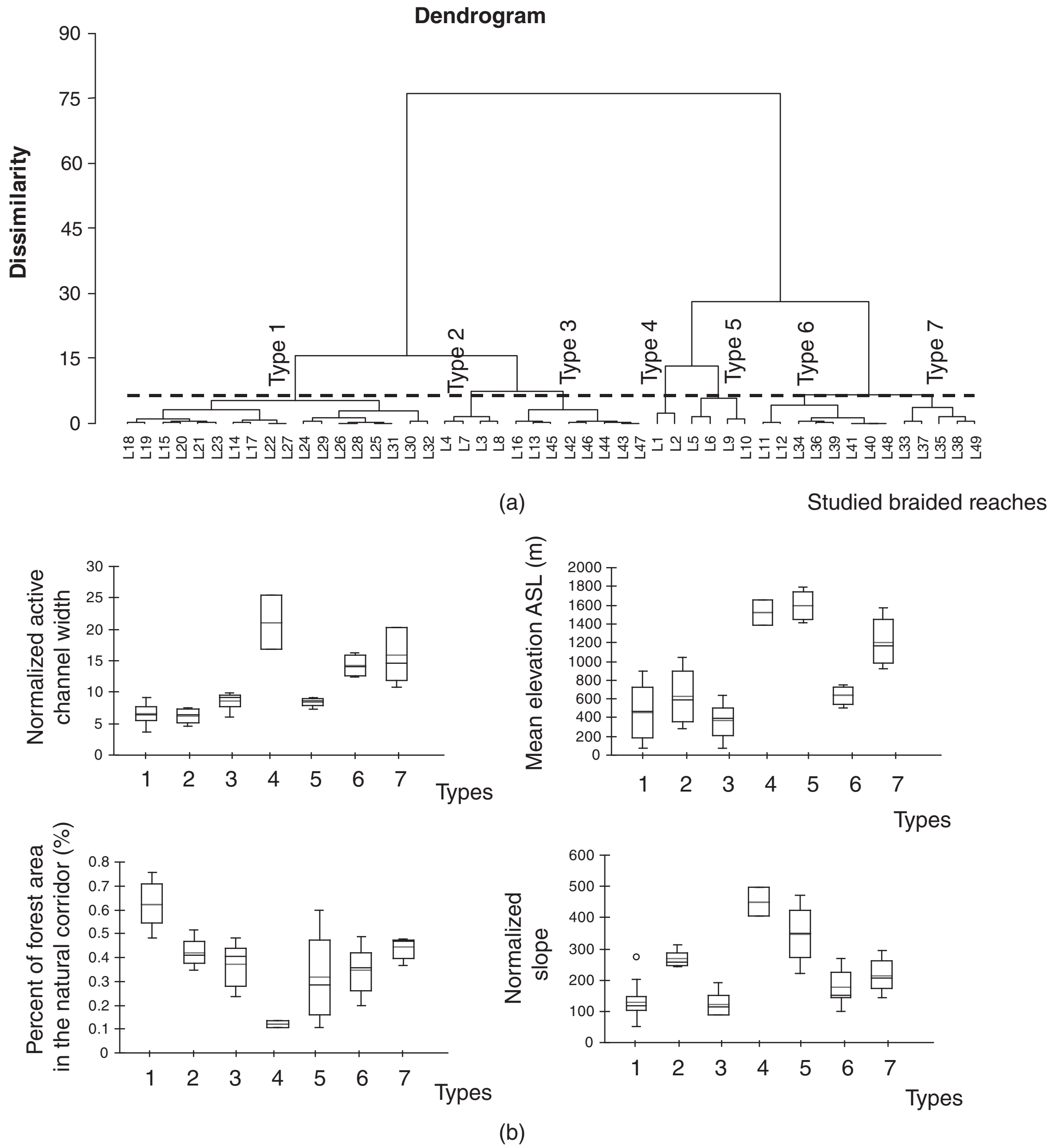
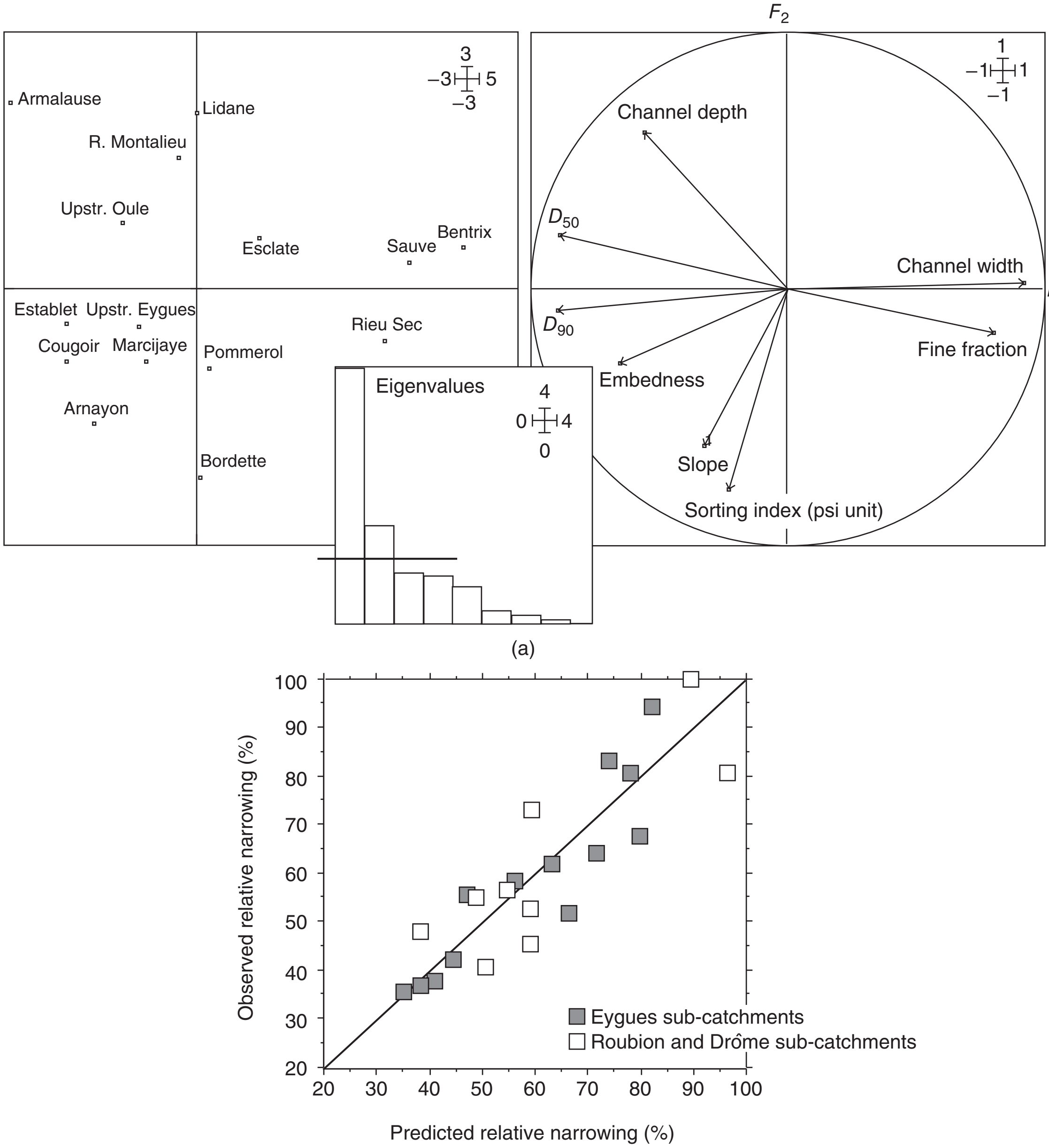


Figure 21.7 Example of cluster analysis performed on a set of 49 braided rivers in the French Alps. (a) Dendrogram highlighting the hierarchy of braided types. (b) Box plots showing the distribution of the different parameters for each of the seven selected braided types: W^* (active channel width rated by the catchment size$^{0.44}$), S^* (slope rated by the catchment size$^{-0.47}$), mean elevation above sea level (ASL), percentage of forest area in natural corridor. Box plots provide the 10th, 25th, 50th, 75th and 90th centiles and the mean (in grey). Source: Piégay *et al*, 2009. Reproduced with permission of Springer.

and ANOVAs) to the cases when the variables to explain are categorical, with two (logistic) or more (multinomial) possible categories. Thus, they help in explaining and predicting the occurrence of an event or another or the individuals' belonging to a group or another). Logistic models are therefore commonly used in medical sciences to distinguish two groups of individuals (e.g. healthy and unhealthy persons, treated and untreated persons). It can be usefully applied in fluvial geomorphology to predict the probability of occurrence of a specific spatial entity according to its human induced or natural characteristics (Table 21.3). Downs (1994 and 1995) used logistic regression to predict the probability of channel adjustment (e.g. widening, shifting, deposition) from continuous (channel slope) and categorical variables (channel environment, channel regulation), describing 285 reaches within the Thames basin (England). He thus built predictive models of the channel response to natural and human controls. For example, the probability of in-channel deposition increases as slope decreases because the



$$\text{Narrowing} = -82.95 + 128.02\,(F_1+5)^{-0.5} + 176.55\,(F_2+5)^{-0.5}$$
$$r^2 = 0.82;\ p = 0.0002$$

(b)

Figure 21.8 Prediction of channel narrowing from channel grain size and geometry characteristics. (a) Projection of the studied streams and variables describing the channel morphology on the first factorial map of a normalized PCA and graph of the eigenvalues. (b) Observations versus predictions of channel narrowing from the first two components of the PCA. The model is performed on the Eygues sub-catchments (grey squares) and validated on the Roubion and Drôme sub-catchments (white squares).

ability of the channel to transport sediment is reduced, but this probability is less in sand/gravel and urbanized basins than in basins with other land uses and substrate types. Silt deposition is the primary sedimentation process in the studied area.

Bledsoe and Watson (2001) used logistic regression to predict thresholds of channel pattern and instability, using the logistic curve as a visual tool to highlight the sensitivity of channel to shifting as a function of specific stream power relative to grain size. Rice (1998) studied a set of torrential basins and assessed their potential bedload supply delivery to the main stem, developing a logistic model to predict the probability that a torrent contributes substantially to the bedload supply

Table 21.3 Explain and predict probabilities through generalized, Bayesian or probabilistic models.

Example of statistical tools	Examples of application in fluvial geomorphology	References
Logistic and multinomial regression	Channel adjustment versus geology/channel gradient/land use/management	Downs (1995)
	Channel pattern versus overbank sediment thickness	Piégay *et al.* (2002)
	Significant tributary in term of sediment source versus relative basin area of the tributary and index of the tributary's sediment delivery potential based on stream power in the tributary	Rice (1998)
	Channel instability versus mobility index slope, discharge, D_{50}	Bledsoe and Watson (2001)
	Occurrence of particle clusters versus reach-scale parameters	Strom and Papanicolaou (2009)
	Debris flow occurrence versus Catchment characters	Bertrand *et al.* (2013)
Neural network	Grain size classes	Krein *et al.* (2008)
Distribution models	Probability of a grain size to be moved/not moved /equilibrium slope	Ferro and Porto (2011)
	Prediction of grain travel distance	Schmidt and Ergenzinger (1992)
	Prediction of velocity distribution	Lamouroux *et al.* (1995)
Bayesian analysis	Probability that elevation changes are true in a DoD map	Wheaton *et al.* (2010)

of the main stem. The explanatory variables were relative basin area and the product of absolute basin area and slope. For the Ain River (France), Piégay *et al.* (2000) showed that the depth of sediment deposited in former channels does not depend on the age of the forms but on their geometry, with formerly meandering channels exhibiting higher sedimentation rates than former braided channels, despite having a younger age. A logistic regression model was applied to predict the probability that a former channel has originated from a meander channel or a braided channel according to its overbank sediment thickness. More recently Bertrand *et al.* (2013) used a logistic regression model to assess the probability for a channel to produce a debris flow according to its basin geometry (e.g. its downstream slope and its Melton index) (Fig. 21.9), which was then used to map likelihood of debris flow occurrence across the entire French Alps.

Explaining and predicting through distributions: distributional modelling and Bayesian analyses

In case the variable of interest is numeric rather than categorical, one reason to consider models explaining and predicting probabilities (rather than, e.g., mean values) is that considering the distribution of the variables of interest rather than a simple summary statistic of their distribution might offer further insight into a phenomenon. The distribution of lengths moved by tagged particles and the duration of particle rest periods was described by exponential or gamma laws (Schmidt and Ergenzinger 1992). For local water velocities, which vary in space and time within reaches, Lamouroux *et al.* (1995) assumed the relative velocity distribution to be a mixture of Gaussian (centred) and exponential (decentred) distributions and modelled the probability density of relative velocity through a maximum likelihood method, with a shape parameter s measuring this mixing. They calculated unexplained variance between predicted and observed frequencies for the velocity classes and found that the best predictors of the shape parameter were determined by using stepwise forward linear regression on averaged variables describing flow conditions (Fig. 21.10).

Probabilities can also be integrated more thoroughly with the models in order, in particular, to illustrate in a non-deterministic way the cascade of causes and consequences at stake in a phenomenon. This can be carried out through the consideration of the phenomenon under study as a random process. Indeed, a random process corresponds to a collection of variables describing states of the system. These might show some cause-to-consequence links, implying, for instance, a certain evolution of the process over time or space. Hence one goal of the use of a random process such as a Markov chain is to assess transition probabilities between states (Tables 21.4). These methods are treated in more detail in Section 21.5.

In Bayesian inference also, the random variables of interest are characterized by their distributions, rather than by point estimates of parameters (such as mean value). Besides, Bayesian models calculate the probability of causes (i.e. distribution of the parameters of interest) based on the probability of consequences (i.e. distribution of observed data) through Bayes' formula. Such a stochastic approach has several advantages. It allows one to incorporate into the results some *a priori* knowledge regarding the variable of interest, and also to deal with the problem of uncertainties, due to e.g. errors in measurement or natural variability. Indeed, in Bayesian analyses, uncertainties are accounted for explicitly and might be propagated along a causal chain. So far, the use of Bayesian statistics remains uncommon in geomorphology. However, it has been used in studies related to wood budgets in streams (Merten *et al.* 2013) and sediment transport, e.g. Schmelter *et al.* (2012) characterized sediment rating curves and their uncertainty through a Bayesian approach, propagating uncertainty into yearly cumulative sediment budgets.

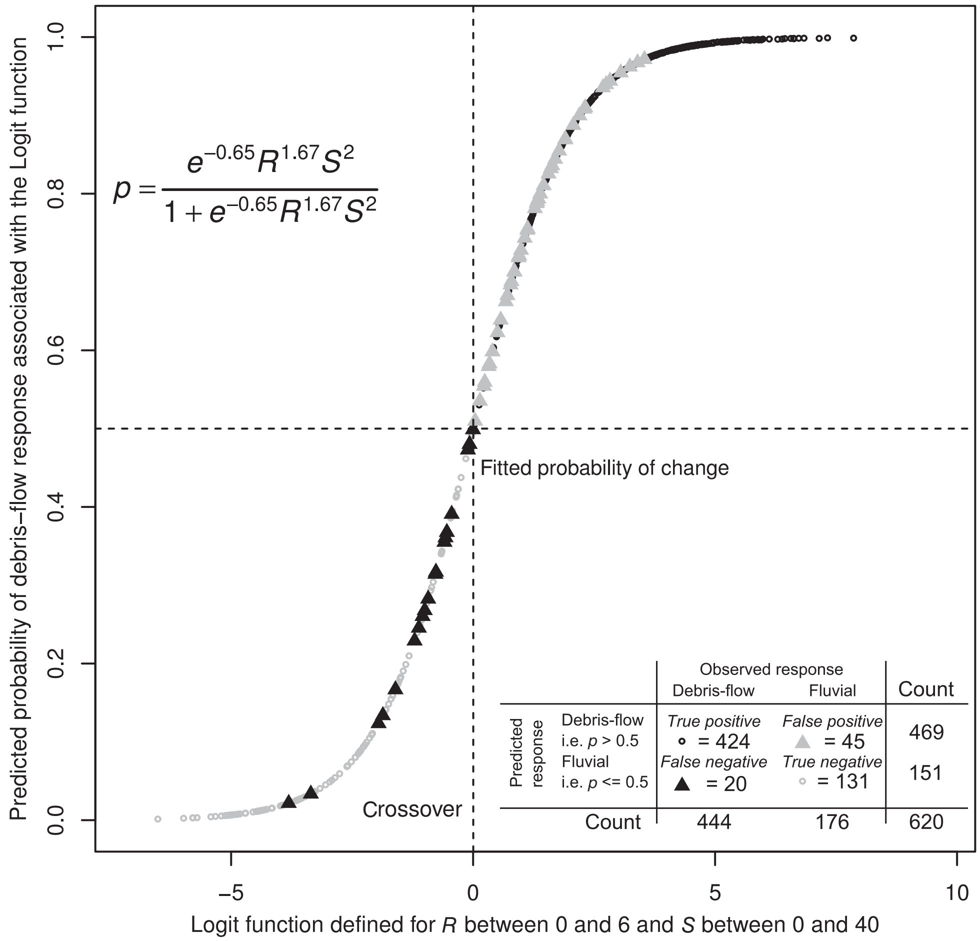


Figure 21.9 Logistic regression model predicting the probability of debris flow occurrence according to the channel slope, *S*, and the Melton index, *R*. Source: Bertrand *et al*, 2013. Reproduced with permission of Springer.

21.5 Describing, explaining and predicting variables in space and time

Spatial and temporal predictors might be treated as standard predictors (e.g. considering a regression of a dependent variable versus time or testing the differences in the mean of a dependent variable according to sites). On the other hand, some methods are specifically designed to deal with spatial and temporal patterns. In particular, they take into account the fact that these patterns might show autocorrelation [i.e. there might be some link between the value of Y at time t and its value at a close time t' or between the value of Y at location (x, y) and its value at a close location (x', y').

Many types of tools specifically exist to describe temporal and spatial patterns, such as fractal analysis, spectral analysis, ARMA (AutoRegressive Moving Average) models, autocorrelation measurements, but also segmentation and associated tests (Tables 21.4). The aim is to evaluate trends, periods, homogeneous segments or some break points in the series and to assess and model the complexity of spatial or temporal information, to assess also time of adjustment or propagation through a given channel length. Such approaches can be conducted with simple statistics such as propagation time fitted on a linear trend (Liébault *et al.* 2005) or model of changes (Demoulin *et al.* 2012).

Analysis of time series has not been widely used in fluvial geomorphology, mostly because long-term series, on which such statistics could be applied to assess thresholds or determine periods, are usually not available. Most of the time series approaches have been developed on hydrological records, notably to assess the flood recurrence interval of the peak flows [see Gordon *et al.* (2004) for an introduction] or the stationarity in long-term hydrological or climatic series.

Analysing spatial and temporal patterns through standard methods

Variables relating to space and time can be treated very well as standard explanatory variables, in particular when they are categorical variables (for instance, relating to different sites or different seasons). Hence regional distinctions can be obtained through comparison of coefficients and exponents in regression equations, as can be done with any kind of grouping of

Table 21.4 Describe, explain and predict data in space and time.

Aim	Example of statistical tools	Examples of application in fluvial geomorphology	References
Describe autocorrelation and periodicity in time and space	Neighbour *K* statistics	Longitudinal pattern of large wood along the channel length	Kraft and Warren (2003)
	Autocorrelation coefficients	Channel width and depth	Robison and Beschta (1990)
		Sediment load, unit-bar abundance and braiding intensity through time	Ashmore (1991)
	Semivariogram, Geary's *C*, Moran's *I*	Bed microprofile statistics	Clifford *et al.* (1992) Madej (1999)
		Automated grain-size measurement	Carbonneau *et al.* (2004) Verdu *et al.* (2005)
		Longitudinal pattern of channel width and vertical evolution	Aubry and Piégay (2001)
	Fractal analysis	Drainage network organization	Gao and Xia (1996)
		Fluvial topography	Pelletier (2007)
		Multiscale statistical properties of a corridor width	Gangodagamage *et al.* (2007)
		Form of particle clusters	Papanicolaou *et al.* (2012)
	Spectral analysis	Channel width and stream gradient	Nakamura and Swanson (1993)
		Bedload transport detection (FFT)	Krein *et al.* (2008)
		Displacement of groups of particles between two sequential images (FFT)–PIV	Hardy *et al.* (2011)
		Meander pattern	Güneralp and Rhoads (2011)
	Wavelet analysis	Hydrological effects of dam	White *et al.* (2005)
		Evolution of sediment load through time	Zhang *et al.* (2008)
	Kriging and other interpolation	Spatial prediction of river channel topography	Legleiter and Kyriakidis (2008)
	Markov chain	Sediment budget	Kelsey *et al.* (1987 Malmon *et al.* (2002)
		Longitudinal succession of in channel features	Grant *et al.* (1990)
		Fluctuating velocity profile	Kirkbride and Ferguson (1995)
Describing and testing breaks in signals	Pettitt and Hubert tests, spatial constraint clustering, hidden Markov model, contrast enhancing	Channel/floodplain width along a hydrog. network	Alber and Piégay (2011) Leviandier *et al.* (2012) Notebaert and Piégay (2013)

the individuals (cf. Section 21.2 on bivariate statistics). This approach is common, for instance, in studies relating watershed area and sediment yields (Poulos *et al.* 1996) where predictive equations are derived from various databases. The weakest point of such an approach is the accuracy of the different methods used to define sediment budgets. Similarly, piedmont sediment accumulations have been linked to their upstream drainage areas, for instance, in alluvial fans from Japan and the southwestern United States (Oguchi and Ohmori 1994). Residual distances can be used to classify observations according to group characteristics within some geographical unit. Alluvial fan types were distinguished in this way in southeast Spain (Silva *et al.* 1992). A two-dimensional plot of high residuals from fan gradient–drainage area and fan area–drainage area was interpreted according to regional knowledge of geology, tectonics and geomorphic evolution of the Guadalentin depression. Large deviations in general channel geometry–discharge relations (Q–width, Q–depth) have been used as indices of local sensitivity to bed modification and in identifying areas where channel design or river restoration is required (Wharton 1995).

Describing autocorrelated patterns and periodicity of signals

Autocorrelation analysis is a way to evaluate periodicity and trends in spatial and temporal data. Autocorrelation can be defined as a similarity between values as a function of their relative position in time or space, such as geometric characteristics along a stream profile. Positive autocorrelation occurs when the values measured on close plots or times are more similar than average.

Spatial autocorrelation functions have not been widely used to describe spatial structures of fluvial forms, but the few examples have mainly addressed the regularity of fluvial facies (pool, riffle) along the long profiles. Statistics such as Moran's I and Geary's c were developed to measure the autocorrelation structure of geomorphological data. Madej (1999) used Moran's

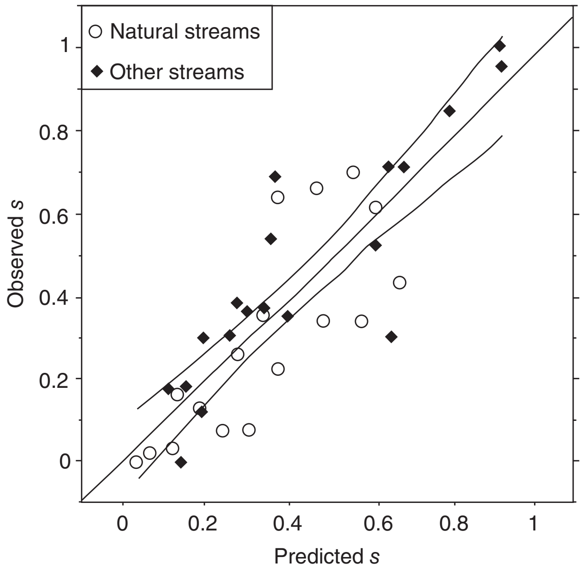


Figure 21.10 Observations versus predictions of *s* from Froude number (*Fr*) and roughness/depth ratio (*D/H*), showing the 95% confidence interval of the linear regression of observations on predictions and the line of perfect agreement. Natural streams are those where neither morphology or discharge were altered by human activities. Source: Lamouroux *et al*, 1995. Reproduced with permission of AGU.

I to detect the presence and scale of significant spatial autocorrelation of bed elevations and also to evaluate the distance above which a bed elevation value is independent of its neighbouring elevations. Clifford *et al.* (1992) used both autocorrelation functions and periodograms to evaluate the geographical scale of roughness elements (e.g. grains and bedforms) and then integrated the results into hydraulic roughness equations to predict mean velocity. Aubry and Piégay (2001) described examples of using spatial autocorrelation functions to describe longitudinal complexity of channel geometry (e.g. trend or repetition of characteristics) and spatial structure of basins (elevation, geological features). We used three sets of simulated data (absence of spatial structure, periodic spatial structure, linear gradient) to compare different autocorrelation functions (Geary's *c*, non-ergodic covariance and correlation). Because each considers the local variance differently, the three functions provide different patterns. In the case of a linear gradient, Geary's *c* grows as a parabolic branch underlining the existing trend, whereas the non-ergodic covariance is bounded and the no- ergodic correlation is zero whatever the distance. As a consequence, the lag distance depends strongly on the statistics used, from 28 km (Geary's *c*) to 10–15 km (non-ergodic covariance). In the case of two-dimensional grid data, omnidirectional analysis can provide an autocorrelation lag which is lower than those provided by one of the directional analyses when the variable is characterized by a geographical orientation. Hence the simultaneous use of the different statistics and the use of more than one direction (rather than a simple omnidirectional analysis) is advisable.

Fractal analysis is another way to study autocorrelation and in particular periodic trends; it describes spatial structure by introducing a scaling perspective. A fractal can be defined as a spatial object comprised of elements that exhibit a similar pattern over all scales. It is possible to define a fractal dimension (D_f) that corresponds to the rate at which the element complexity changes with the scale. Fractal analysis in fluvial geomorphology has mainly focused on drainage networks (Gao and Xia 1996). Amongst the 13 papers dealing with fractal analysis published in *Water Resources Research* between 1987 and 1997, 11 concerned drainage networks. In another application of these techniques, Nestler and Sutton (2000) used fractals to characterize cross-sectional distributions of area and energy as a function of scale to evaluate effects of river regulation on aquatic habitat. For a cross-section of the Missouri River, they plotted an energy–area graph showing the modification of historical habitats by regulation works. Under intermediate flows (906 $m^3 s^{-1}$), the existing conditions no longer contain large-scale habitat components (oval pictograms) that were present in the past.

Repetitions of patterns can also be assessed through other signal processing methods. One example of such an analysis is the use of the finite Fourier transforms (Hardisty 1993). Time series and spatial continua can be viewed as a sum of basic sinusoids of varying amplitudes and frequencies. Fourier transforms decompose the energy (or variability) of such signals into the energies at all possible periods (or frequencies), hence providing periodograms that describe the power spectrum of the signal. While Fourier analysis characterizes signals in the frequency domain and is unable to account for changes in patterns in the time (or space) domain, wavelets account for variabilities in the signal on a time–frequency or space–frequency domain. A signal can therefore be decomposed into its changes at different temporal or spatial scales. Such a decomposition might be used with several purposes, in particular to filter a signal. Most of the time, wavelets have been used to characterize time series such as discharge to identify different temporal patterns (e.g. seasonal periodicity, changes in peak flows). Following these approaches in hydrology, Keylock *et al.* (2014) recently used wavelets to explore flow hydraulics, focusing on bedform flow velocity-intermittency structures of bedform flows. Contributions to the characterization of longitudinal geomorphic patterns are still rare. For instance, Lashermes *et al.* (2007) first used wavelet decomposition to describe changes in a topographic signal (specifically, in local curvatures and slope direction changes) at various scales, then thresholded them to keep only meaningful changes, which were then re-aggregated to extract a stream network. We used wavelet decomposition to analyse variations in the talweg elevation of the Rhône River to explore downstream organization of in-channel features (e.g. sequence of pools and riffles). Segmenting the wavelet coefficients series through Whitcher's segmentation method (Whitcher *et al.* 1999) mainly showed shifting variability levels on a scale of 0.6 km (Fig. 21.11). Local variation of the

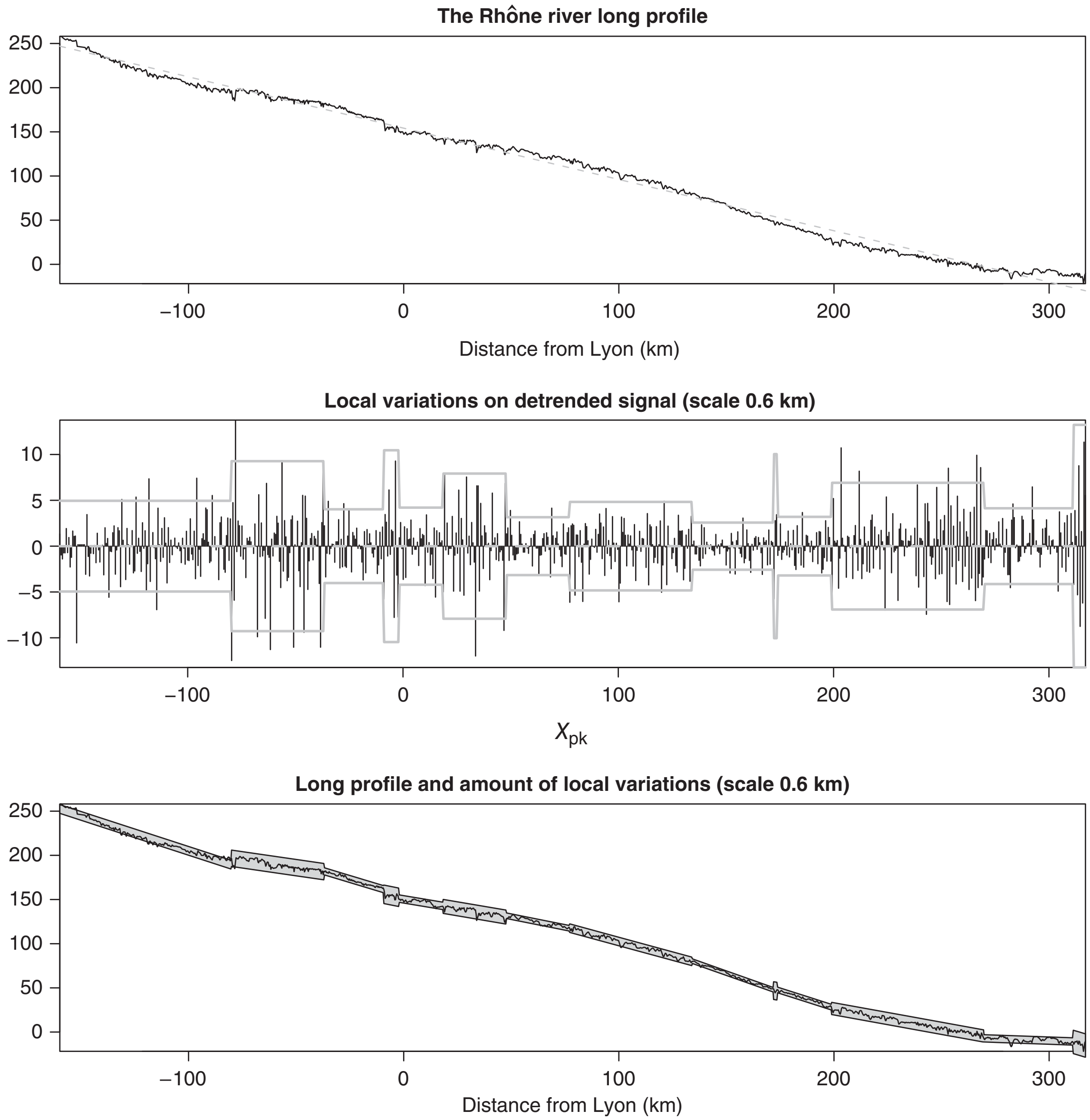


Figure 21.11 Local variations (scale 0.6 km) of the Rhône river talweg elevation, as estimated through wavelet decomposition and Whitcher's segmentation.

detrended signal (e.g, standard deviations of wavelet coefficients) at the 0.6 km scale (the most detailed one) showed significant differences amongst well-identified reaches. Over 500 km, in-channel features of the Rhône River evince different patterns. Some reaches have frequent forms with large amplitudes (lowland rivers with deep pools exacerbated by active mining), whereas others have smaller features (upland reaches characterized by mobile bars and active sediment transport or flat bedrock-floored reaches). Comparing current patterns with those of the early 20th century, deep pools are much more common now than previously and the Rhône is losing its characterics of an Alpine upland river. The longitudinal organization is not related to the position of the 19 dams along the Rhône, indicating that the channel adjustments occurred in response to embankments in the late 19th and early 20th century, rather than dams.

Describing, modelling and predicting the evolution of variables in space and time

Random processes such as Markov chains can be used, for example, in fluvial sediment transport to assess transition probabilities determined by flood events and storage in reaches. Probabilities of particle transitions from one 'storage reservoir' to others were derived from computed residence times in a stream of northern California by Kelsey *et al.* (1987). Applications included estimating flushing times of sediment out of a reach and changes in sediment masses stored in the four sediment storage reservoirs. This approach yields better results with a longer hydrological data record (in this case 35 years) and with rapid rates of morphological changes in the channel, as in this example.

The Markov chain approach has also been used to highlight the longitudinal distribution of channel units (e.g. pools, riffles, rapids, cascades, steps) along stream reaches, indicating that channel units occur in non-random, two-unit sequences (Grant *et al.* 1990). Two-unit sequences have been described by a matrix of transition probability where each cell is the probability with which a given morphological unit is followed downstream by another given morphological unit (e.g. a pool followed by a riffle). Preferred sequences, defined as the sequences occurring more frequently than expected from random sequence, demonstrated that steps, cascades and rapids are frequently followed by pools in Lookout Creek, Oregon. The sequences are slightly different in the steeper French Pete Creek, Oregon, where the cascade–pool and rapid–pool sequences are unfrequent, but riffle–cascade sequences are common, reflecting the higher gradient and supply of large boulders from debris flows (Fig. 21.12).

The question of the nature of flow turbulence has been tackled through different statistical approaches: either turbulence is regarded as a spatially independent and temporally random phenomenon, thus without memory, or as a structurally coherent phenomenon. In the former case, descriptive parameters such as standard deviations of velocity or shear stress can be used, but this approach would not be appropriate if interactions between layers, internal structures and periodic behaviour are considered, because transitions are inherited from previous states. Temporal fluctuations in streamwise and vertical velocities at different depths over a gravel bed can be analysed in terms of a Markov process, once the studied variables (e.g. horizontal

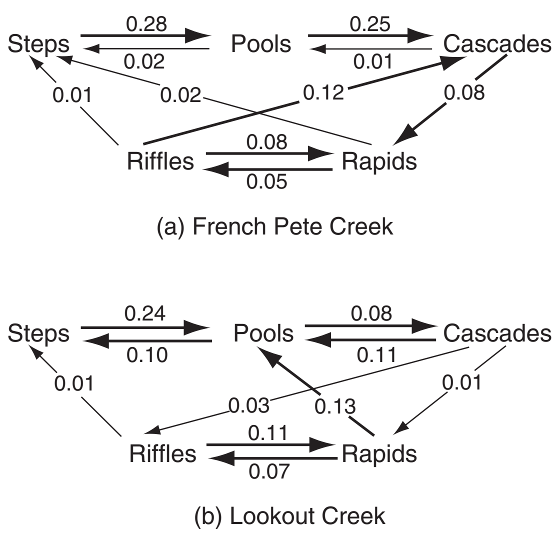


Figure 21.12 Sequence relation diagrams for French Pete Creek (a) and Lookout Creek (b) (Western Cascades, Oregon). Numbers shown are differences between observed and random transitional probabilities. Bolder arrows indicate transitional probabilities >0.05. Source: Grant *et al*, 1990. Reproduced with permission of Geological Society of America.

velocity or directions of vertical movement) have been transformed into categorical variables. Statistical properties of the Markov chain are then tested against the null hypothesis of absence of spatial structure. This hypothesis is invalidated by the demonstration of some more frequent states and transitions (Kirkbride and Ferguson 1995).

This approach is convenient when processes are partly understood and where interdependence of variables makes the definition of functional links difficult. Randomness is assumed in a series of states or events and the probabilities of a change from a state to another one are estimated. These probabilities account for the length of sequences, such as (in sedimentology) simulation of depositional units in space and time.

Describing and testing breaks in series

Other approaches can be used to focus on point breaks in spatial and temporal continua, thus identifying distinct homogeneous segments (Brunel 2000). A break in a temporal series can be defined as a change in the distribution of the series at a given time *t*. Different tests or statistics, such as the Lee and Heghinian test (Lee and Heghinian 1977), the Pettitt test (Pettitt 1979) or Buishand's *U* statistic (Buishand 1984), can identify such breaks. For example, the Pettitt test is a non-parametric test based on the Mann–Whitney test, with the null hypothesis being the absence of a break in the series X_i of size N. For all times t, the statistic $U_{t,N}$ considers the two time series $(X_1, X_2, \ldots, X_t)$ and $(X_{t+1}, X_{t+2}, \ldots, X_N)$ are drawn from the same distribution. The segmentation test (Hubert 1989) is another way first to describe non-stationary series by detecting several breaks and then testing the simultaneous significance of all breaks between adjacent segments. The operator can define the number of segments required and their minimal size, or an optimization algorithm can be used to identify the best segmentation amongst all possible ones.

Studies of the non-stationarity of discharge series can be used to detect changes occurring during the last century due to human or climate modifications within catchments. The gauging station at Luc-en-Diois (Drôme, France) has recorded daily discharge (Q_d) since 1907. The series of annual peak flows was studied for possible breaks and homogeneous time sequences using Buishand, Lee and Heghinian and Pettitt tests and using the Hubert segmentation algorithm. The segmentation was performed for two, three, four and five segments, each of which was required to include at least five events. For the annual peak flow series, Pettitt's test identified a possible change in the 1930s, which was unclear according to other statistics. When using a variable describing the form of morphogenic peak flow events ($Q_{di} > Q_{d1.5}$), such as the residuals of the linear relationship between the flood water volume and the flood duration, most of the stationarity tests used (mainly Hubert, Pettit, Lee and Heghinian) validated a break in the trend in the 1930s (Fig. 21.13). Hence different variables describing peak flows and different segmentation tests pointed to a statistically valid hydrological change around the 1930s. The change was

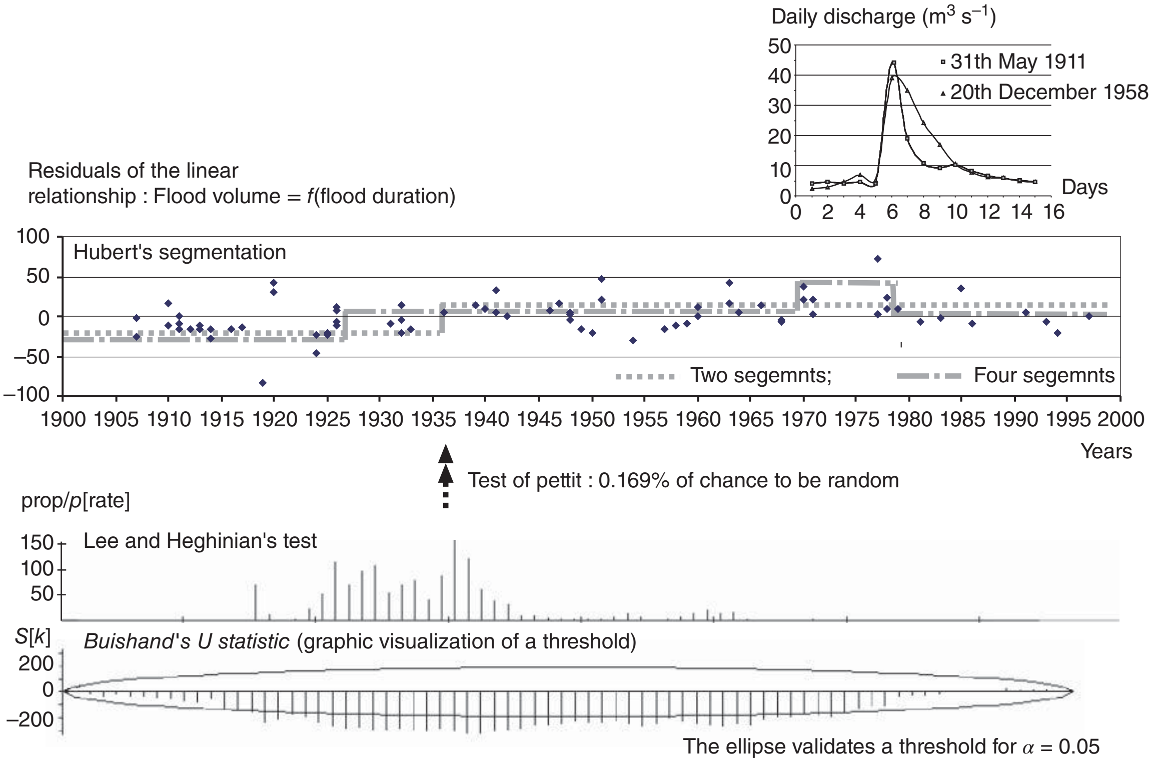


Figure 21.13 Threshold tests and Hubert's segmentation performed on a response variable related to the daily discharge of the Luc-en-Diois gauging station (1907–1997). The response variable is the residuals of the relationships between the flood volume and the flood duration. Each selected flood event is higher than $Q_{1.5}$ in recurrency and flood duration is calculated using a threshold discharge of 10 m^3 s^{-1} as base level.

less pronounced in magnitude than in the shape of the flood hydrographs. Floods were flashier, with sharper peaks before the 1930s than after, consistent with the hydrological response expected from catchment afforestation and cessation of grazing, which occurred two to three decades before.

When studying spatial structure, in particular when the spatial structure can be reduced to just one dimension, segmentation and segmentation tests can be carried out. We applied the Hubert segmentation to the first component of a PCA summarizing variables describing the longitudinal pattern of the channel shifting of the Willamette River, Oregon. Maps of the channel reach (from Eugene to the confluence with the Columbia River) on four dates (1850, 1895, 1932 and 1995) were cut into 1 km long sections. The channel maps were overlayed using a GIS system and we extracted channel change variables (e.g. channel narrowing, eroded floodplain area, constructed floodplain area) during the three periods (1850–1895, 1895–1932 and 1932–1995). The segmentation analysis on the first factorial axis defined homogeneous reaches in terms of channel shifting (e.g. the spatial structure) whatever the temporal trend (Fig. 21.14). When using the segmentation in four segments (statistically validated on the Wald threshold $\alpha = 0.05$), from km 17 to km 80 (Saint Paul), the river was characterized by a very stable channel, whereas from km 81 to km 100 (near Salem), and also from km 150 to km 223 (downstream of Eugene), the river channel is highly mobile whatever the period concerned. Between km 100 and km 150, the pattern is slightly more contrasted longitudinally but much less mobile than in its two neighbouring segments. The second factorial axis provides the spatio-temporal changes: downstream from Salem, no real change occurred in channel shifting, whereas the channel underwent narrowing from 1895 to 1932 between km 110 (Salem) and km 160 (Albany) and to km 200 (Monroe). More recently (1932–1995), the channel narrowed in the reach from km 110 to km 160.

21.6 Relevance and limitations of statistical tools

Tables 21.5 and 21.6 summarize the general advantages and constraints of statistical tools in fluvial geomorphology. With increasing amounts of information (considerably increased by automatic recording and environmental data banks), classified objects must be compared in both space and time (e.g. sediment yields in different basins and at different flows). Differences and ordering must be statistically validated before any interpretation.

The choice of tool from the broad array available should be informed by the aim of the study and the availability, type and

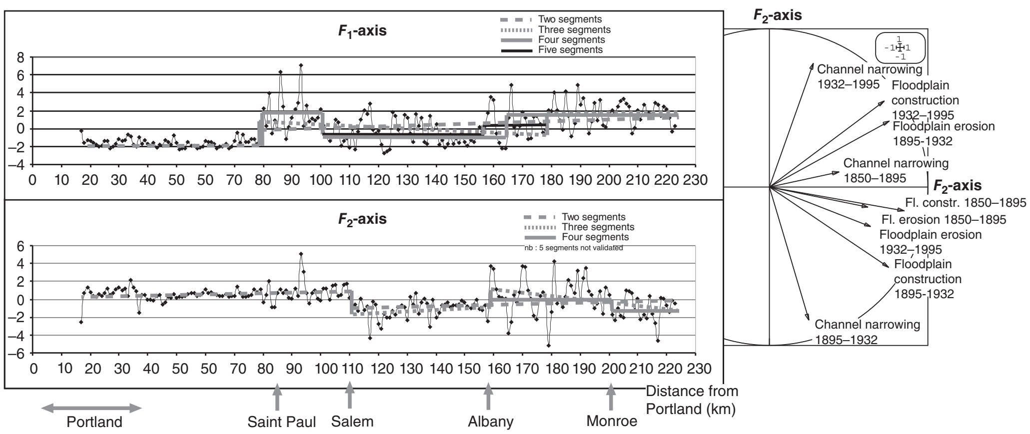


Figure 21.14 Hubert segmentation performed on the longitudinal continuum of the Willamette River between its confluence with the Columbia and Eugene (223 km). Each elementary channel segment is 1 km long and described from four different maps. A previous data table of 207 slices × 9 variables describing channel planform changes has been summerized by a PCA into two main factors, which are then segmented using the segmentation test of Hubert.

Table 21.5 Relevance and advantages of statistical tools.

Pedagogy. Statistics and graphic extensions can highlight phenomena, at first sight unclear, because of simplification and ordering capability. It is very useful for making comparisons between geomorphic forms and events
Objectivity. Statistics permit reduction of subjectivity and expand discussion of results regarding other controlling factors
Flexibility. The number of statistical tools is so large that it is always possible to find one that might help you to interpret your information if it is numerical. Moreover, it is possible to create specific statistics for a given question. With the increase in computer capacity, randomization tests considerably enhance the statistical relevance
Prediction. Statistics could produce models that evaluate the sensitivity of fluvial systems to disturbance
Multiple-scale capability. Statistics can consider elements/objects at different temporal and spatial scales and can also examine interaction of scales. Moreover, statistics permit large-scale, holistic approaches
Interdisciplinary. Statistics are widely used in environmental sciences (geology, geography, ecology, hydrology) and are consequently a means for interdisciplinary collaboration, which is currently a key concept for success in applied research

Table 21.6 Constraints and limitations of statistical tools.

In the formulation of scientific hypothesis. The use of statistics may impose a specific framework for sampling, data collection, interpretation and validation of hypotheses. Is the scientific question confirmed by statistical tools or do the tools guide the scientific question?
Assumption and accuracy of tools. Standard statistical tools are based on preliminary hypotheses (e.g. linearity, homoscedasticity, independence, normality). Some tests are robust, others are more sensitive concerning the maintenance of these assumptions. There are some constraints to using statistical tools when there are gaps into the data, or when we obtain a mixture of categorical and continuous data
Results. The quality of the results is not dependent of whether or not one uses statistics but on the quality and originality of the geomorphological question one poses and solves
Prediction. Models are empirical, and not necessarily related to the physical laws that control the geomorphological processes. As a consequence, they are often limited in context and use. Numerical models that are based on physical laws may need many assumptions and simplifications to be accurate, which also limits their use to a particular functional context
Psychology. The use of statistics may be suspect. Where is the verification? It is often said that statistics can demonstrate all and can hide the poorness of the results and the data

degree of confidence in the data. Selecting the optimal sampling strategy should be based on factors such as randomness, representativeness, method standardization, feasibility and cost. Most of the time, it is not possible to consider an entire population or an entire area. For example, the number of grains of sand and gravel in a bar, the number of different velocity values that could be recorded on a channel cross-sections and the spatial variability of many variables over large areas are too great. Sampling strategies are needed to extract efficiently and cost-effectively a representative sample that accurately reflects the characteristics of the population or the area. When the entire population or area is known, probability sampling can be used, including random, systematic, stratified or clustered procedures. However, in fluvial geomorphology, it is often difficult to know the distribution of a population and thereby predetermine the sample size necessary to obtain an estimate of given precision. Scaled stratification within geographical information systems is often useful for sampling.

Quantifying precision and uncertainty when measuring

One of the main issues in using data and then applying statistics is data reliability, due to possible errors in measurements. 'What is the error of detection when measuring ^{137}Cs or ^{210}Pb activity?' is an important question when testing for differences between sites or samples. 'What is the RMS (root mean square) error when georectifying satellite images or air photographs?' must be addressed when overlapping images and reaching conclusions about channel changes. The question 'Is there any bias in the measure due to techniques/protocols used or operators?' must be addressed to determine whether relationships among different datasets are geomorphically meaningful. We can also use statistics to calibrate and establish corrections to allow two different measures of the same object or process to be compared.

ANOVA, chi-squared tests or t-tests are the tools most commonly used to assess whether the measurements are biased by operators or by methods. For example, Thévenet *et al.* (1998) measured geometric volumes of wood jams composed of both wood pieces and air as the product width × height × length. Before determining a linear model linking wood mass and air–wood volume, it was necessary to confirm that there was no estimator bias. Measurements were made by three operators and an ANOVA test was performed to test the independence between them. The null hypothesis was 'no significant difference between the three operators'. The null hypothesis was accepted at a confidence level $1 - \alpha = 95\%$ (i.e. there was less than a 5% chance of rejecting it if it was true), validating the procedure of field data measurement (degrees of freedom = 2; $F = 0.661$; $p = 0.52$).

Field sampling of sediment in gravel-bed rivers has been widely debated, especially since the introduction of the pebble count technique by Wolman (1954). Issues include the best method of sampling (bulk versus sieving) and operator bias. Wohl *et al.* (1996) characterized the variability among replicates of a sampling method, among four methods and among operators. Three types of tests were used: t-tests to evaluate differences in the D_{50} for each operator (e.g. veteran, experienced and novice), ANOVA to determine whether any of the methods yield statistically different distribution parameter estimates (D_{50} and D_{84}) and chi squared test to evaluate differences in phi class distributions. Three of the four methods produced values of D_{50} and D_{84} that were not statistically different. However, grain-size distributions by different operators yielded samples that were statistically different (D_{50}, D_{84}, distribution of size classes, variance). Inevitably, measurement errors are not always precisely or equally known for all the included variables. Indeed, they are generally inseparable from other sources of errors, e.g. unknown environmental factors, such as in regressions that associate discharge with width, depth and velocity (Rhoads 1992).

Improving confidence interval when sampling

Along with measurement errors, all natural systems show a certain heterogeneity linked to unknown environmental factors. Indeed, there might be some unexplained variability among individuals due to, for example, temporal or spatial patterns. For instance, variations in sediment supply rates may reflect the passage of bed forms and groups of particles. Sampling should last long enough to take short-term fluctuations into account, and when using point samples, sampling should be dense enough to characterize variations across the streams (Gomez and Church 1989).

The sampling strategy should be designed to address at least partly these problems. For instance, spatial autocorrelation functions are useful for calibrating a sampling design when it concerns a geographical area. Indeed, a common hypothesis in many statistical techniques is the independence of the data values collected. In this context, it is important to determine the lag of the spatial dependence, i.e. the distance above which the data values are independent of one another. If we know the lag, we can define a grid sampling design in which the grid width exceeds the lag, and we can then use the classical statistic tests on the sample to infer the results to the entire population. We did so before developing a logistic regression model to assess the probability of occurrence of gullies in the Roubion basin, south-eastern France, according to various geographical parameters (e.g. slope, altitude, land use). A sample of pixels was extracted from GIS covers in order to assess the spatial autocorrelation of the variables controlling the gully occurrence. The question was to determine a lag above which the altitude or slope values could be considered independent of one another. Geary's c test was performed for these two variables for 70 lag classes with a step of 50 m each (Fig. 21.15). In order to assess the lag of the positive spatial autocorrelation, some p-values were calculated from randomization tests (1000 random permutations of the two values z_i and z_j for each class) and were plotted as a joined function of the autocorrelation function (Fig. 21.15). For a given threshold (e.g. $\alpha = 0.05$), the values were then distinguished as being significantly autocorrelated or not and this confidence interval was then plotted on the graph of the c of Geary function (Fig. 21.15). Once we observed a sharp change between a sequence of low p-values and a sequence of high p-values, we considered that the lag had been reached. In this case, the lag was reached over 2450–2500 m for the altitude and 2550–2600 m for the slope. In order to respect the assumption of independence in the modelling process, systematic sampling should be conducted within a grid where each sample should be separated from the others by 2600 m.

Another important aspect of sampling strategy is to determine the minimum number of samples that one has to collect so that the results can be inferred to the whole populations under study. This is the focus of power analysis. Indeed, statistical power relates to the ability of a test to detect significant links between variables. Power analysis is relatively easy to carry out when one knows *a priori* that the variable of interest follows (approximately) a standard distribution. It is much less straightforward in the other cases. Although power analysis has, to our knowledge, never been applied in geomorphological studies, it is gaining interest in other areas of research [see, for instance, Mozayyan *et al.* (2011) for a hydrological example].

In a more general way, the definitions of optimum sampling strategies rely on estimates of uncertainty (e.g. standard deviation from the mean) and in particular on the calculation of confidence intervals for the parameters of interest. One topic that has inspired a considerable literature about sampling methods is the measurement of size of coarse-grained sediments. Because we do not know the mean and the standard deviation of the grain size of the gravel population, but also because its distribution does not follow the normal law, we cannot determine with accuracy the best sampling size. Resampling procedures, such as bootstrap simulation techniques, can be helpful in determining the best sample; this method produces confidence intervals for the parameter of interest without requiring any distributional assumptions. Such techniques were used by Rice and Church (1996a) to determine percentile standard errors in Wolman counts so as to evaluate the sample size needed to maximize the precision of grain size estimated within a gravel-bed river. At each of two sites, they measured the b-axis of around 3500 particles, applying the procedure for 20 runs from $n = 50$ to $n = 1000$ and repeated 200 times to estimate a standard error for D_5, D_{16}, D_{25}, D_{50}, D_{75}, D_{84} and D_{95}. They also calculated the theoretical normal percentile standard errors and compared them with bootstrap percentile standard errors (Fig. 21.16). They obtained two main results: (i) while D_{50} standard errors were consistently low, fine-tail percentile errors were underestimated by the normal model and coarse-tail percentile errors were overestimated, demonstrating that for a given precision it would be necessary to collect more and less particles, respectively, than that expected by a normal distribution to characterize the distribution tails; (ii) for sample sizes exceeding 300–400 particles, the marginal gains in precision were small relative to the additional sampling effort.

Validation of explanatory models

The residual error of the models might be due to measurement errors but also to unknown or unaccounted for factors. Introduction of new or more adapted explanatory variables in the model is then performed if possible. However, the more complex the statistical tools, the larger number of data they require for statistical validity. Moreover, in some studies, causality or even correlation is difficult to establish because of multicolinearity: one or several other explanatory variables are correlated with each other and with the response. Such is the case for the link between channel width and meander wavelength, both of

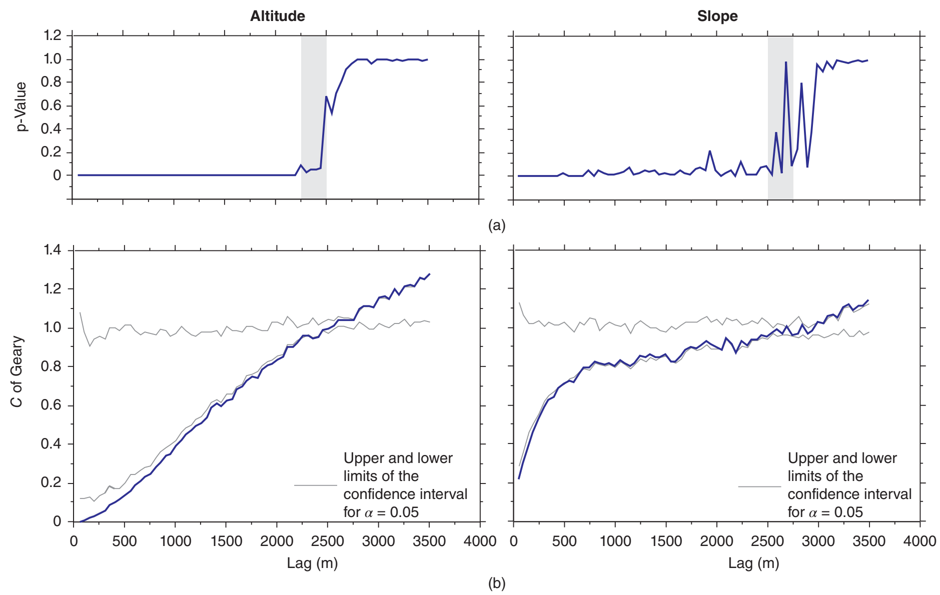


Figure 21.15 Determination of the threshold above which the values are not autocorrelated: example of elevations and slopes of the Bine and Soubrion catchments (France) (samples of 50 × 50 m pixels extracted on a digital elevation model). (a) *p*-Values of randomization test (1000 random permutations of the neighbouring values at each 50 m step). When the *p*-value averages 0, the positive autocorrelation is high. (b) Geary's *C* and the confidence interval under the null hypothesis at $\alpha = 0.05$.

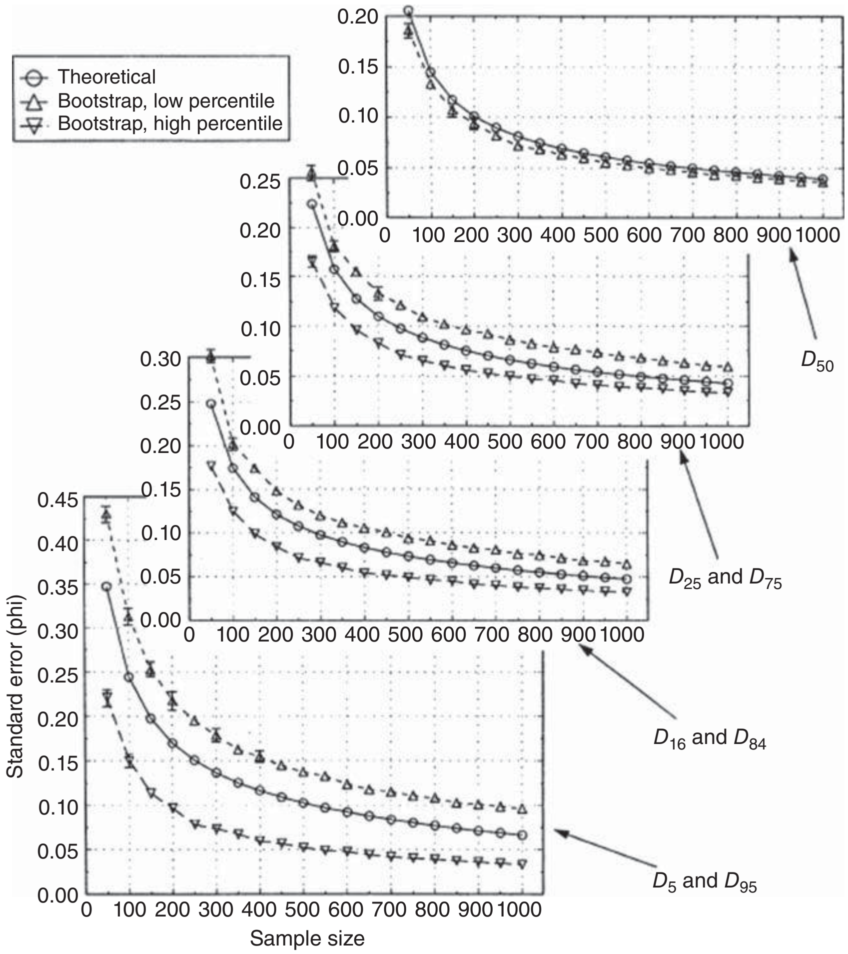


Figure 21.16 Comparison of bootstrap and normal percentile standard errors (Mamquam River site, British Columbia). Error bars indicate the 95% confidence intervals about the mean bootstrap results based on 10 replications. Error bars are shown only where they exceed symbol dimensions. From Rice and Church (1996a), reproduced with permission of SEPM Society for Sedimentary Geology).

which depend on charateristics of the flow dynamics (Chorley and Kennedy 1971). Multiple regressions should be used only for defining response equations in which the net influence of regressors on interacting variables is described (Hey 1978). Preliminary multivariate analyses on a local case study can indicate which are the most significant and predictive variables before extending the area of research. However, the choice may be refuted because of regional heterogeneity.

Relationships between variables sometimes describe physical laws in process–response systems, often hydraulic relations. If these relationships can be described *a priori*, functional or structural analysis should be used instead of least-squares regression (Mark and Church 1977). In morphological system analysis, however, the knowledge we can acquire on such relationships is, necessarily empirical. As such, it is valid only for defined geographical areas or the value range of the regressor variables (Hey and Thorne 1983). Further, establishing causality and correlations is particularly difficult when dealing with a process that is not at an equilibrium (Ebisemiju 1988). Mass equilibrium is more rapidly attained in small watersheds (Church and Mark 1980), which implies that catchment comparisons should consider size effects. Strong correlations could be interpreted as evidence of such an equilibrium. However, complexity arises from unequal responses to process changes in the various subsystems so that deviations could be clues either to inherited features or to different, sometimes unsuspected, behaviour.

Validation of underlying hypotheses

Classical inferential statistics (e.g. classical linear regressions) often rely on hypotheses such as a Gaussian distribution and homoskedasticity of residuals. Statistical and graphical residual analysis is recommended in order to verify these requirements. A transformation of the data is actually often carried out prior to fitting the model. Geometric progressions in datasets, such as length, area and slope angle, entail logarithmic, square root or dimensionless normalizations

Some extensions of these models have been implemented to make such hypotheses unnecessary and/or to allow for a wider variety of situations. Generalized linear models, for instance, apply to data such that the model's residuals are distributed according to laws such as binomial, multinomial, Poisson or negative binomial.

In the cases where inferential tests do not exist or cannot be applied, other procedures involving Monte Carlo simulations and whose recent development has been possible because of increases in computer capacity can be applied. For instance, permutational and randomization tests are more flexible in term of assumptions than classical tests and also provide p-values. Because the distribution of the statistics under H_0 is built with some data and not some samples, the tests concern only the data and not a population from which these data could be inferred. The goal of the permutation test is to generate all the possible values of a given statistic (all the permutations of the values amongst the individuals) in order to calculate the p-value associated with the observed value of the statistics. When the censing of all the permutations is not possible ($n > 10$), such tests can be approximated by a randomization test that is based on a limited number of permutations rather than doing all of the possible ones (Manly 1991).

Predictive performance of statistical models

Statistics are used not only as explanatory tools but also for providing predictions. For example, it is possible to use the multiple regression model of Fig. 21.8 to predict the intensity of channel changes (e.g. narrowing) according to basic field characteristics such as its grain size characteristics and geometry. We can use the logistic model of Fig. 21.9 to predict the probability of debris flow occurrence according to the channel slope (S) and the Melton index (R) at an entire catchment scale (e.g. ~19,000 km of upland stream length of the southern French Alps). Prediction from regression analysis requires critical analysis, to avoid potential errors such as generalizing predictions to inappropriate conditions, outside the relevant geographical area or range of variables.

To assess the extent to which the results of a statistical analysis will generalize to an independent dataset, different procedures can be considered. Cross-validation, for instance, requires separating the dataset into two: one on which the model is fitted (its parameters are estimated) and one on which its ability to fit another dataset is validated. There are no clear rules in terms of sharing the dataset, 1/3 and 2/3, 1/4 and 3/4 or half-and-half are common. Following this, the predictive performance (for the second part of the data) of the model (fitted on the first part of the data) can be assessed through, e.g., a confusion matrix, displaying the rate of true positives (a), false positives (b), false negatives (c) and true negatives (d). From the model sensitivity, $a/(a + c)$, specificity, $d/(b + d)$, and overall accuracy, $(a + d)/(a + b + c + d)$, it may be possible to measure both the user's and producer's accuracies. In particular, the producer's accurary is used to estimate the model capacity to classify accurately the different categories compared with all the field observations. Following Bertrand *et al.* (2013), the global sensitivity and specificity for the 620 models are 0.95 and 0.75, respectively, and the percentage of correct classification is 0.89.

Alternatively to cross-validation, when the number of observations is fairly low, one may prefer to use the leave-one-out procedure. This consists in fitting n models each accounting for $n - 1$ individuals and then validating them on the individual left out. Observed versus predicted values might then be displayed to assess the predictive quality of the model. Such an approach was applied by Bertrand *et al.* (2013), who performed a leave-one-out validation and calculated the 95% confidence intervals for the logistic regression coefficients (Fig. 20.9). Performances based on such validation showed the 95% confidence intervals of β_0, $\beta_1(R)$ and $\beta_2(S)$ to be −0.60 and 0.70, 1.63 and 1.70, and 1.98 and 2.03, respectively.

21.7 Conclusion

Statistics is a universal tool whose language is understood by scientists whatever their discipline. By allowing the scientist to interpret environmental data with high temporal and spatial variability, statistics complement physical- and experiment-based methods, facilitating the interpretation of fluvial forms and processes in their diversity (regional, longitudinal, size, time). Indeed, variability is not only due to measurement imprecision or experimental error, but is a fundamental attribute of environmental data, which requires specific approaches for assessment and understanding. Statistical tools can support causal research and predictions and help provide an experimental framework where hypotheses can be formulated, tested and validated, allowing laws and then theories to be produced.

Application of statistical tools in fluvial geomorphology has the advantages of reducing subjectivity, eliminating assumptions, facilitating comparison between different spatial and temporal datasets of large sizes, refining data collection, revealing exceptions or new relations, predicting performance and improving system analysis. In a systems approach, statistical tools have become more applicable through the increasing expansion of computational capacity and increasingly convenient statistical software. Given the mixed nature of datasets increasingly used by fluvial geomorphologists to understand process and form, the more sophisticated statistical tools can offer benefits to research in the field. To develop more realistic descriptions of fluvial morphological systems, process–response systems, time and space trends and size effects, will require the collection of sufficient data and more thought about their relevance. Appropriate statistical analyses can contribute to the interpretation of these data.

However, to date, statistical tools have not been used nearly to their full potential in fluvial geomorphology. Statistics are only one type of tool among many and cannot solve every geomorphological question. To date, statistical applications in fluvial geomorphology have been dominated by linear regression, justly criticized as a 'black-box' empirical approach. Another criticism of statistics concerns the naivity of many interpretations and the seeming possibility of concluding everything and its contrary. When conducting these studies, errors can be made at the stage of experimental design and choice of the statistics used, usually tracable to misunderstanding of the purpose and limitations of the statistical analyses. Other errors occur when interpreting the results, notably in causal interpretation when the researchers ask the data to say what they cannot say.

Acknowledgements

We gratefully thank Pierre Clément who was a co-author of the first edition of this chapter. L. Ashkenas, M. Bertrand, S. Dufour, S. Gregory, N. Landon, F. Liébault, E. Parrot and S. Tacon contributed to the data collections used to exemplify the present discussions.

References

Alber, A. and Piégay, H. 2011. Spatial disaggregation and aggregation procedures for characterizing fluvial features at the network-scale: application to the Rhône basin (France). *Geomorphology* 125(3): 343–360.

Ashmore, P. 1991. Channel morphology and bed load pulses in braided, gravel-bed streams. *Géografiska Annaler* 73A: 37–52.

Aubry, P. and Piégay, H. 2001. Pratique de l'analyse de l'autocorrélation spatiale en géomorphologie fluviale: définitions opératoires and test. *Géographie Physique and Quaternaire* 55(2): 115–133.

Batalla, R.J. 1997. Evaluating bed-material transport equations using field measurements in a sandy gravel-bed stream, Arbucies River, NE Spain. *Earth Surface Processes and Landforms* 22: 121–130.

Batalla, R.J. and Sala, M. 1995. Effective discharge for bedload transport in a subhumid Mediterranean sandy gravel-bed river (Arbucies, NE Spain). In: Hickin, E.J., ed., *River Geomorphology*, Chichester: John Wiley & Sons, pp. 93–103.

Belletti, B., Dufour, S. and Piégay, H. 2013. Regional variability of aquatic pattern in braided reaches (example of the French Rhône basin). *Hydrobiologia* 712: 25–41.

Bertrand, M., Liébault, F. and Piégay, H. 2013. Debris-flow susceptibility of upland catchments. *Natural Hazards* 67: 497–511.

Bledsoe, B.P. and Watson, C.C. 2001. Logistic analysis of channel pattern thresholds : meandering, braiding and incising. *Geomorphology* 38: 281–300.

Bombino, G., Gurnell, A.M., Tamburino, V., Zema, D.A. and Zimbone, S.M. 2009. Adjustments in channel form, sediment calibre and vegetation around check-dams in the headwater reaches of mountain torrents, Calabria, Italy. *Earth Surface Processes and Landforms* 34: 1011–1021.

Bravard, J.P. and Petit, F. 1997. *Les Cours d'Eau. Dynamique du Système Fluvial*, Paris: Armand Colin, 222pp.

Brookes, A., 1987. River channel adjustments downstream from channelization works in England and Wales. *Earth Surface Processes and Landforms* 12: 337–351.

Brunel, S. 2000. *Utilisation de l'Hydrologie Quantifiée pour l'Étude des Changements Fluviaux and de l'Éventuelle Discrimination de Leurs Causes Climatiques and Anthropiques dans le Cas de 10 Cours d'Eau Français*, unpublished DEA Interface Nature Société, Université de Lyon 3 and Cemagref, 157pp.

Buishand, T.A. 1984. Tests for detecting shift in the mean of hydrological time series. *Journal of Hydrology* 73: 51–69.

Carbonneau, P., Lane, S. and Bergeron, N. 2004. Catchment-scale mapping of surface grain size in gravel bed rivers using airborne digital imagery. *Water Resources Research* 40(7): W07202.

Chessel, D. and Mercier, P. 1993. Couplage de triplets statistiques and liaisons espèces-environnement. In: Lebreton, J.D. and Asselain, B., eds., *Biométrie and Environnement*, Paris: Masson, pp. 15–43.

Chorley, R.J. and Kennedy, B.A. 1971. *Physical Geography. A Systems Approach*, London: Prentice Hall, 370pp.

Church, M. and Mark, D.M. 1980. On size and scale in geomorphology. *Progress in Physical Geography* 4: 342–390.

Clément, P. and Piégay, H. 2003. Statistics and fluvial geomorphology. In: Kondolf, G.M. and Piégay, H., eds., *Tools in Fluvial Geomorphology*, Chichester: John Wiley & Sons, pp. 597–630.

Clifford, N.J., Robert, A. and Richards, K.S. 1992. Estimation of flow resistance in gravel-bedded rivers: a physical explanation of the multiplier of roughness length. *Earth Surface Processes and Landforms* 17: 111–126.

Constantine, J.A., Dunne, T., Piégay, H. and Kondolf, G.M. 2010. Controls on the alluviation of oxbow lakes by bed-material load along the Sacramento River, California. *Sedimentology* 57(2): 389–407.

Dawson, M. 1988. Sediment size variation in a braided reach of the Sunwapta river, Alberta, Canada. *Earth Surface Processes and Landforms* 13: 599–618.

Demoulin, A., Beckers, A. and Bovy, B. 2012. On different types of adjustment usable to calculate the parameters of the stream power law. *Geomorphology* 138: 203–208.

Downs, P.W. 1994. Characterisation of river channel adjustments in the Thames basin, south-east England. *Regulated Rivers – Research and Management* 9: 151–175.

Downs, P.W. 1995. Estimating the probability of river channel adjustment. *Earth Surface Processes and Landforms* 20: 687–705.

Ebisemiju, F.S. 1988. Canonical correlation analysis in geomorphology with particular reference to drainage basin characteristics. *Geomorphology* 1: 331–342.

Ebisemiju, F.S. and Ekiti, A. 1989. A morphometric approach to gully analysis. *Zeitschrift für Geomorphologie* 33: 307–322.

Emery, J.C., Gurnell, A.M., Clifford, N.J., Petts, G.E., Morrissey, I.P.an Soar, P.J. 2003. Classifying the hydraulic performance of riffle–pool bedforms for habitat assessment and river rehabilitation design. *River Research and Applications* 19: 533–549.

Ferguson, R.I. 1986a. Hydraulics and hydraulic geometry. *Progress in Physical Geography* 10: 1–31.

Ferguson, R.I. 1986b. River loads underestimated by rating curves. *Water Resources Research* 22: 74–76.

Ferguson, R.I. 2008. Gravel-bed rivers at the reach scale. In: Habersack, H., Piégay, H. and Rinaldi, M., eds., *Gravel-bed Rivers VI: from Process Understanding to River Restoration*, Amsterdam: Elsevier, pp.33–60.

Ferguson, R.I., Cudden, J.R., Hoey, T.B. and Rice, S.P. 2006. River system discontinuities due to lateral inputs: generic styles and controls. *Earth Surface Processes and Landforms* 31: 1149–1166.

Ferro, V. and Porto, P. 2011. Predicting the equilibrium bed slope in natural streams using a stochastic model for incipient sediment motion. *Earth Surface Processes and Landforms* 36(8): 1007–1022.

Fonstad, M.A. and Marcus, W.A. 2010. High resolution, basin extent observations and implications for understanding river form and process. *Earth Surface Processes and Landforms* 35(6): 680–698.

Gao, J. and Xia, Z. 1996. Fractals in physical geography. *Progress in Physical Geography* 20: 178–191.

Gangodagamage, C., Barnes, E., and Foufoula-Georgiou, E. 2007. Scaling in river corridor widths depicts organization in valley morphology. *Geomorphology* 91(3–4): 198–215.

Gomez, B. and Church, M. 1989. An assessment of bedload sediment transport formulae for gravel bed rivers. *Water Resources Research* 25: 1161–1186.

Gordon, N.D., MacMahon, T.A., Finlayson, B.L., Gippel, C.J. and Nathan, R.J. 2004. *Stream Hydrology, an Introduction for Ecologists*, 2nd edn., Chichester: John Wiley & Sons, 429pp.

Grant, G.E., Swanson, F.J. and Wolman, M.G. 1990. Pattern and origin of stepped-bed morphology in high-gradient streams, Western Cascades, Oregon. *Geological Society of America Bulletin* 102: 340–352.

Gregory, K.J. and Walling, D.E. 1973. *Drainage Basin Form and Process. A Geomorphological Approach*, London: Edward Arnold, 458pp.

Güneralp, I. and Rhoads, B.L. 2011. Influence of floodplain erosional heterogeneity on planform complexity of meandering rivers. *Geophysical Research Letters* 38(14): L14401.

Gurnell, A.M. 1997. Adjustments in river channel geometry associated with hydraulic discontinuities across the fluvial–tidal transition of a regulated river. *Earth Surface Processes and Landforms* 22: 967–985.

Gurnell, A., Goodson, J., Thompson, K., Clifford, N. and Armitage, P. 2007. The river-bed: a dynamic store for plant propagules? *Earth Surface Processes and Landforms* 32(8): 1257–1272.

Hardisty, J. 1993. Time series analysis using spectral techniques: oscillatory currents. *Earth Surface Processes and Landforms* 18: 855–862.

Hardy, R.J., Best, J.L., Parsons, D.R. and Keevil, G.M. 2011. On determining the geometric and kinematic characteristics of coherent flow structures over a gravel bed: a new approach using combined PLIF–PIV. *Earth Surface Processes and Landforms* 36: 279–284.

Hassan, M.A., Church, M. and Ashworth, P.J. 1992. Virtual rate and mean distance of travel of individual clasts in gravel-bed channels. *Earth Surface Processes and Landforms* 17: 617–627.

Hatfield, R.G. and Maher, B.A. 2009. Fingerprinting upland sediment sources: particle size-specific magnetic linkages between soils, lake sediments and suspended sediments. *Earth Surface Processes and Landforms* 34(10): 1359–1373.

Hey, R.D. 1978. Determinate hydraulic geometry of river channels. *Journal of the Hydraulics Division, American Society of Civil Engineers* 104(HY6): 869–885.

Hey, R.D. and Thorne, C.R. 1983. Hydraulic geometry of gravel-bed rivers. In: *Proceedings, 2nd International Symposium on River Sedimentology, Nanjing*, pp. 713–723.

Hoey, T. 1992. Temporal variations in bedload transport rates and sediment storage in gravel-bed rivers. *Progress in Physical Geography* 16: 319–338.

Horton, R.E. 1945. Erosional development of streams and their drainage basin; hydrophysical approach to quantitative morphology. *Geological Society of America Bulletin* 56: 275–370.

Howard, A.D. and Hemberger, A.T. 1991. Multivariate characterization of meandering. *Geomorphology* 4: 161–186.

Hubert, P. 1989. Segmentation des séries hydrométéorologiques: application à des séries de précipitations et de débits de l'Afrique de l'Ouest. *Journal of Hydrology* 110: 349–367.

Kelsey, H.M., Lamberson, R. and Madej, M.A. 1987. Stochastic model for the long-term transport of stored sediment in a river channel. Water Resources Research 23: 1738–1750.

Keylock, C.J., Singh, A., Venditti, J.G. and Foufoula-Georgiou, E., 2014. Robust classification for the joint velocity–intermittency structure of turbulent flow over fixed and mobile bedforms. *Earth Surface Processes and Landforms* 39: 1717–1728.

Kirkbride, A.D. and Ferguson, R. 1995. Turbulent flow structure in a gravel-bed river: Markov chain analysis of the fluctuating velocity profile. *Earth Surface Processes and Landforms* 20: 721–733.

Knighton, A.D., Woodroffe, C.D. and Mills, K. 1992. The evolution of tidal creek networks, Mary River, Northern Australia. *Earth Surface Processes and Landforms* 17: 167–190.

Kraft, C.E. and Warren, D.R. 2003. Development of spatial pattern in large woody debris and debris dams in streams. *Geomorphology* 51: 127–139.

Krein, A., Klinck, H., Eiden, M., Symader, W., Bierl, R. Hoffmann, L. and Pfister, L. 2008. Investigating the transport dynamics and the properties of bedload material with a hydro-acoustic measuring system. *Earth Surface Processes and Landforms* 33(1): 152–163.

Kunhle, R.A. 1992. Bed load transport during rising and falling stages on two small streams. *Earth Surface Processes and Landforms* 17: 191–197.

Lamouroux, N., Souchon, Y. and Herouin, E. 1995. Predicting velocity frequency distributions in stream reaches. *Water Resources Research* 31: 2367–2375.

Lashermes, B., Foufoula-Georgiou, E. and Dietrich W.E. 2007, Channel network extraction from high resolution topography using wavelets. *Geophysical Research Letters* 34: L23S04.

Lecce, S.A. 1997. Spatial patterns of historical overbank sedimentation and floodplain evolution, Blue River, Wisconsin. *Geomorphology* 18: 265–277.

Lee, A.F.S. and Heghinian, S.M. 1977. A shift of the mean level in a sequence of independent normal random variables: a Bayesian approach. *Technometrics* 19(4): 503–506.

Legleiter, C.J. and Kyriakidis, P.C. 2008. Spatial prediction of river channel topography by kriging. *Earth Surface Processes and Landforms* 33(6):841–867.

Leopold, L.B. and Maddock, T. 1953. *The Hydraulic Geometry of Streams Channels and Some Physiographic Implications*, US Geological Survey Professional Paper 252, Washington, DC: US Government Printing Office, 57pp.

Leopold, L.B. and Wolman, M.G. 1957. *River Channel Patterns; Braided, Meandering and Straight*, US Geological Survey Professional Paper 282-B, Washington, DC: US Government Printing Office, pp. 39–85.

Leopold, L.B., Wolman, M.G. and Miller, J.P. 1964. *Fluvial Processes in Geomorphology*, San Francisco, CA: W.H. Freeman.

Leviandier, T., Alber, A., Le Ber, F. and Piégay, H. 2012. Comparison of statistical algorithms for detecting homogeneous river reaches along a longitudinal continuum. *Geomorphology* 138(1): 130–144.

Liébault, F., Clément, P., Piégay, H., Rogers, C.F., Kondolf, G.M. and Landon, N. 2002. Contemporary channel changes in the Eygues basin, southern French Prealps: causes of regional variability according to watershed characteristics. *Geomorphology* 45: 53–66.

Liébault, F., Gomez, B., Page, M., Marden, M., Peacock, D., Richard, D. and Trotter, C.M. 2005. Land-use change, sediment production and channel response in upland regions. *River Research and Applications* 21: 739–756.

Liébault, F., Lallias-Tacon, S., Cassel, M. and Talaska, N. 2013. Long profile responses of Alpine braided rivers in SE France. *River Research and Applications* 29(10): 1253–1266.

Llorens, P., Querault, I., Plama, F. and Gallart, F. 1997. Studying solute and particulate sediment transfer in a small Mediterranean mountainous catchment subject to land abandonment. *Earth Surface Processes and Landforms* 22: 1027–1035.

Madej, M.A. 1999. Temporal and spatial variability in thalweg profiles of a gravel-bed river. *Earth Surface Processes and Landforms* 24: 1153–1169.

Malmon, D.V., Dunne, T. and Reneau, S.L. 2002. Predicting the fate of sediment and pollutants in river floodplains. *Environmental Science and Technology* 36: 2026–2032.

Manly, B.F.J. 1991. *Randomization and Monte Carlo Methods in Biology*, London: Chapman and Hall, 281pp.

Mark, D.M. and Church, M. 1977. On the misuse of regression in earth science. *Mathematical Geology* 9: 63–75.

Mazeika, S., Sullivan, P., Watzin, M.C. and Hession, W.C. 2004. Understanding stream geomorphic state in relation to ecological integrity: evidence using habitat assessments and macroinvertebrates. *Environmental Management* 34(5): 669–683.

Merten, E.C.,Vaz, P.G., Decker-Frite, J.A., Finlay, J.C. and Heinz, S.G. 2013. Relative importance of breakage and decay as processes depleting large wood from streams. *Geomorphology* 190: 40–47.

Mozayyan, M., Akhondali, A. and Basiri, R. 2011 Determination of optimum sample size in regression analysis for some hydrologic variables with emphasis on power analysis. *International Journal of Applied Environmental Sciences* 2: 217–227.

Nakamura, F. and Swanson, F.J. 1993. Effects of coarse woody debris on morphology and sediment storage of a mountain stream system in western Oregon. *Earth Surface Processes and Landforms* 18: 43–61.

Navratil, O., Breil, P., Schmitt, L., Grosprêtre, L. and Albert, M.B. 2013. Hydrogeomorphic adjustments of stream channels disturbed by urban runoff (Yzeron River basin, France). *Journal of Hydrology* 485(2): 24–36.

Nelson, P.A., Bellugi, D. and Dietrich, W.E. 2014. Delineation of river bed-surface patches by clustering high-resolution spatial grain size data. *Geomorphology* 205: 102–119.

Nestler, J. and Sutton, V.K. 2000. Describing scales of features in river channels using fractal geometry concepts. *Regulated Rivers – Research and Management* 16: 1–22.

Notebaert, B. and Piégay, H. 2013. Multi-scale factors controlling the pattern of floodplain width at a network scale: the case of the Rhône basin, France. *Geomorphology* 200: 155–171.

Oguchi, T. and Ohmori, H. 1994. Analysis of relationships, among alluvial fan area, source basin area, basin slope, sediment yield. *Zeitschrift für Geomorphologie* 38: 405–420.

Owens, P.N., Walling, D.E. and Leeks, G.J. 1999. Use of floodplain sediment cores to investigate recent historical changes in overbank sedimentation rates and sediment sources in the catchment of the River Ouse, Yorkshire, UK. *Catena* 36: 21–47.

Papanicolaou, A.T., Tsakiris, A.G. and Strom, K.B. 2012. The use of fractals to quantify the morphology of cluster microforms. *Geomorphology* 139: 91–108.

Park, J. 1992. Suspended sediment transport in a mountainous catchment. *Scientific Reports, Institute of Geoscience, University of Tsukuba, Section A* 13: 138–190.

Parsons, M. and Thoms, M.C. 2007. Hierarchical patterns of physical–biological associations in river ecosystems. *Geomorphology* 89: 127–146.

Passmore, D.G. and Marcklin, M.G. 1994. Provenance of fine-grained alluvium and late Holocene land-use change in the Tyne basin, northern England. *Geomorphology* 9: 127–142.

Pelletier, J. D. 2007. Fractal behavior in space and time in a simplified model of fluvial landform evolution. *Geomorphology* 91(3–4): 291–301.

Petit, F. 1987. The relationship between shear stress and the shaping of the bed of a pebble-loaded river, La Rulles, Ardennes. *Catena* 14: 453–468.

Pettitt, A.N. 1979. A non-parametric approach to the change-point problem. *Applied Statistics* 28(2): 126–135.

Piégay, H., Bornette, G., Citterio, A., Hérouin, E., Moulin, B. and Statiotis, C. 2000. Channel instability as control factor of silting dynamics and vegetation pattern within perifluvial aquatic zones. *Hydrological Processes* 14(16–17): 3011–3029.

Piégay, H., Bornette, G. and Grante, P. 2002. Assessment of silting-up dynamics of 11 cut-off channel plugs on a free-meandering river (the Ain River), France. In: Allison, R.J., ed., *Applied Geomorphology, Theory and Practice*, Chichester: John Wiley & Sons, pp. 227–247.

Piégay, H., Alber, A., Slater, L. and Bourdin, L. 2009. Census and typology of braided rivers in the French Alps. *Aquatic Sciences* 71(3): 371–388.

Piégay, H., Kondolf, M., Minear, T., Vaudor, L. 2015. Trends in publications in fluvial geomorphology over two decades: A truly new era in the discipline owing to recent technological revolution?, *Geomorphology* 248: 489–500.

Poulos, S.E., Collins, M. and Evans, G. 1996. Water-sediment fluxes of Greek rivers, southeastern Alpine Europe: annual yields, seasonal variability, delta formation and human impact. *Zeitschrift für Geomorphologie* 40: 243–261.

Reid, I. and Frostick, L.E. 1986. Dynamics of bedload transport in Turkey Brook, a coarse-grained alluvial channel. *Earth Surface Processes and Landforms* 11: 143–155.

Reneau, S.L. and Dietrich, W.E. 1991. Erosion rates in the southern Oregon coast range: evidence for an equilibrium between hillslope erosion and sediment yield. *Earth Surface Processes and Landforms* 16: 307–322.

Rhoads, B.L. 1992. Statistical models of fluvial systems. *Geomorphology* 5: 433–455.

Rhoads, B.L. and Miller, M.V. 1991. Impact of flow variability on the morphology of a low-energy meandering river. *Earth Surface Processes and Landforms* 16: 357–368.

Rice, S. 1998. Which tributaries disrupt downstream fining along gravel-bed rivers? *Geomorphology* 22: 39–56.

Rice, S. and Church, M. 1996a. Sampling surficial gravels: the precision of size distribution percentile estimates. *Journal of Sedimentary Research* 66(3): 654–665.

Rice, S. and Church, M. 1996b. Bed material texture in low order streams on the Queen Charlotte Islands, British Columbia. *Earth Surface Processes and Landforms* 21: 1–18.

Richards, K. 1976. The morphology of riffle–pool sequence. *Earth Surface Processes and Landforms* 1: 71–88.

Rickenmann, D. 1997. Sediment transport in Swiss torrents. *Earth Surface Processes and Landforms* 22: 937–951.

Riquier, J., Piégay, H. and Michalková, M.S. 2015. Hydromorphological conditions in eighteen restored floodplain channels of a large river: linking patterns to processes. *Freshwater Biology* 60: 1085–1103.

Ridenour, G.S. and Giardino, J. 1995. Discriminant function analysis of computational data: an example from hydraulic geometry. *Physical Geography* 15: 481–492.

Robison, E.G. and Beschta, R.L. 1990. Coarse woody debris and channel morphology interactions for undisturbed streams in southeast Alaska. *Earth Surface Processes and Landforms* 15: 149–156.

Schmelter, M.L., Erwin, S.O. and P.R. Wilcock, 2012. Accounting for uncertainty in cumulative sediment transport using Bayesian statistics. *Geomorphology* 175–176: 1–13.

Schmidt, K.H. and Ergenzinger, P. 1992. Bedload entrainment, travel lengths, step lengths, rest periods – studied with passive (iron, magnetic) and active (radio) tracer techniques. *Earth Surface Processes and Landforms* 17: 147–165.

Schumm, S.A. 1960. *The Shape of Alluvial Channels in Relation to Sediment Type*, US Geological Survey Professional Paper 352-B, Washington, DC: US Government Printing Office, pp.17–30.

Sear, D.A., 1996. Sediment transport processes in pool–riffle sequences. *Earth Surface Processes and Landforms* 21: 241–262.

Shields, N.D. 1936. *Anwendung der Ähnlichkeitsmechanik und der Turbulenzforschung auf die Geschiebelerwegung*, PhD Thesis, Berlin: Mitteilungen der Preussischen Versuchanstalt fur Wasserbau und Schiffbau der Universität Berlin, Heft 26.

Silva, P.G., Harvey, A.M., Zazo, C. and Goy, J.L. 1992. Geomorphology, depositional style and morphometric relationships of Quaternary alluvial fans in the Guadalantin Depression (Murcia, Southeast Spain). *Zeitschrift für Geomorphologie* 36: 325–341.

Strahler, A.N. 1952. Dynamic basis of geomorphology. *Geological Society of America Bulletin* 63: 923–938.

Strahler, A.N. 1954. Statistical analysis in geomorphic research. *Journal of Geology* 62: 1–25.

Strom, K.B. and Papanicolaou, A.N. 2009. Occurrence of cluster microforms in mountain rivers. *Earth Surface Processes and Landforms* 34: 88–98.

Tacon, S., Liébault, F. and Piégay, H. 2014. Step by step error assessment in braided river sediment budget using airborne LiDAR data. *Geomorphology* 214: 307–323.

Thévenet, A., Citterio, A. and Piégay, H. 1998. A new methodology for the assessment of large woody debris accumulations on highly modified rivers (example of two French piedmont rivers). *Regulated Rivers – Research and Management* 14: 467–483.

Thoms, M.C., Parsons, M.E. and Foster, J.M. 2007. The use of multivariate statistics to elucidate patterns of floodplain sedimentation at different spatial scales' *Earth Surface Processes and Landforms* 32(5): 672–686

Tucker, L.R. 1958. An inter-battery method of factor analysis. *Psychometrika* 23: 111–136.

Verdu, J.M., Batalla, R.J. and Martinez-Casasnovas, J.A. 2005. High-resolution grain-size characterisation of gravel bars using imagery analysis and geo-statistics. *Geomorphology* 72: 73–93.

Walling, D.E. and He, Q. 1998. The spatial variability of overbank sedimentation on river floodplains. *Geomorphology* 24: 209–223.

Wharton, G. 1995. The channel-geometry method: guidelines and applications. *Earth Surface Processes and Landforms* 20: 649–660.

Wheaton, J. M., Brasington, J., Darby, S.E. and Sear, D.A. 2010. Accounting for uncertainty in DEMs from repeat topographic surveys: improved sediment budgets. *Earth Surface Processes and Landforms* 35(2): 136–156.

Whitcher, B., Guttorp, P. and Percival, D.B. 1999. Multiscale detection and location of multiple variance changes in the presence of long memory, *Journal of Statistical Computation and Simulation* 68: 65–88.

White, M.A., Schmidt, J.C. and Topping, D.J. 2005. Application of wavelet analysis for monitoring the hydrologic effects of dam operation: Glen Canyon Dam and the Colorado River at Lees Ferry, Arizona. *River Research and Applications* 21(5): 551–565.

Wiederkehr, E., Belletti, B., Dufour, S. and Piégay, H. 2010. Physical characterisation of river corridors from orthophotos: challenging issues and first application to the Rhône hydrographic network. In: Addink, E.A. and Van Coillie, F.M.B., eds., *GEOBIA 2010 – Geographic Object-Based Image Analysis, Ghent University, Belgium, 29 June–2 July*, ISPRS Archives, vol. XXXVIII-4/C7.

Wilcock, P.R. 1993. Critical shear stress of natural sediments. *Journal of Hydraulic Engineering* 119: 491–505.

Wohl, E. and Merritt, D. 2005. Prediction of mountain stream morphology. *Water Resources Research* 41: W08419.

Wohl, E.E., Antony, D.J., Masden, S.W. and Thompson, D.M. 1996. A comparison of surface sampling methods for coarse fluvial sediments. *Water Resources Research* 32: 3219–3226.

Wohl, E., Kuzma, J.N. and Brown, N.E. 2004. Reach-scale channel geometry of a mountain river. *Earth Surface Processes and Landforms* 29(8): 969–981.

Wolman, M.G. 1954. A method of sampling coarse river bed material. *Transactions, American Geophysical Union* 35: 951–956.

Wolman, M.G. and Miller, R.J.P. 1960. Magnitude and frequency of forces in geomorphic processes. *Journal of Geology* 68: 54–74.

Woodsmith, R.D. and Buffington, J. 1996. Multivariate geomorphic analysis of forest streams: implications for assessment of land use impacts on channel condition. *Earth Surface Processes and Landforms* 21: 377–394.

Yu, L. and Oldfield, F. 1993. Quantitative sediment source ascription using magnetic measurements in a reservoir-catchment system near Nijar, S.E. Spain. *Earth Surface Processes and Landforms* 18: 441–454.

Zhang, Q., Chen, G., Su, B., Disse, M., Jiang, T. and Xu, C.-Y. 2008. Periodicity of sediment load and runoff in the Yangtze River basin and possible impacts of climatic changes and human activities. *Hydrological Sciences Journal* 53(2): 457–465.

SECTION VII

Conclusion: Applying the Tools

CHAPTER 22

Integrating geomorphological tools to address practical problems in river management and restoration

Hervé Piégay[1], G. Mathias Kondolf[2] and David A. Sear[3]
[1]Université de Lyon, UMR 5600 CNRS, Lyon, France
[2]University of California, Berkeley, CA, USA
[3]University of Southampton, Southampton, UK

22.1 Introduction

Fluvial geomorphology can be useful to other scientists such as ecologists than those involved in this topics (e.g. to provide a framework within which to analyse habitats), and engineers, and also practitioners, such as planners and river managers (e.g. to understand risks and effects of flooding, to regulate in-channel mining or navigation) and those who implement ecological conservation or restoration programmes (e.g. through insights into the physical functioning of ecosystems and constraints posed by human alterations).

Geomorphological questions posed by other scientists and practitioners are often complex and merit being subdivided into a set of more specific questions. The physical, chemical and biological interactions in river systems operate at multiple temporal and spatial scales (Fig. 22.1), which implies that to understand relations or to solve problems will typically require the application of multiple tools. Some of these tools are proper to geomorphology, whereas others were developed in allied fields (such as biology or engineering sciences) and are applied to geomorphological problems. These tools range widely in the temporal and spatial scales of application, from a few minutes or hours (the duration of the bedload movement during a flood event) to several centuries (the time needed for a fluvial system to adjust its geometry to a climate change) and from centimetres (benthic invertebrate habitat) to thousands of square kilometres (large river catchments).

Through the range of tools presented in this book, we have sought to provide a reference not only for the practicing geomorphologists and graduate students, but also for the managers and scientists trained in other disciplines who work with geomorphologists, to understand better the range of approaches potentially available to address problems associated with fluvial forms and processes. This chapter provides a framework within which the tools can be used and presents examples of the application of geomorphic tools to problems in river management and restoration.

22.2 Motivations for applying fluvial geomorphology

> It should be possible to persuade decision-makers that incorporating historical or empirical geomorphic information into river management strategies is at least as valuable as basing decisions on precise, yet fallible, mechanistic models.
>
> *Rhoads (1994)*

This statement captures the sense of potential for applied fluvial geomorphology that rose in concert with a growth in environmental awareness and political will to recognize and account for the environment in land and water management. Since the late 1980s, applied fluvial geomorphology has raised the operational and policy agendas of river management authorities, most recently propelled by the demands for 'morphological' assessment in support of river restoration (Sear *et al.* 1995; Sear *et al.*, 2010; and Gurnell *et al.* 2016).

With increasing emphasis on environmental river management and interest in sustainable approaches to the use of water (and other natural) resources, managers must base their decisions on insights from a variety of disciplines. Because fluvial geomorphology provides the overall framework within which habitats develop, ecological processes operate, floods propagate and waters may (or may not) undergo purification en route to the river and downstream, geomorphological analyses are central to understanding many issues in river management, including maintenance and restoration of aquatic and riparian habitats, flood risk and water quality. Specifically, fluvial geomorphologists are increasingly called upon to answer questions at temporal and spatial scales different from those which other disciplines have typically employed. Graf (1996) described this resurgence of geomorphological application as the 'return to its roots of a close association with environmental resource management and public policy', arguing that geomorphology is now mature enough, after a period characterized by a focus

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

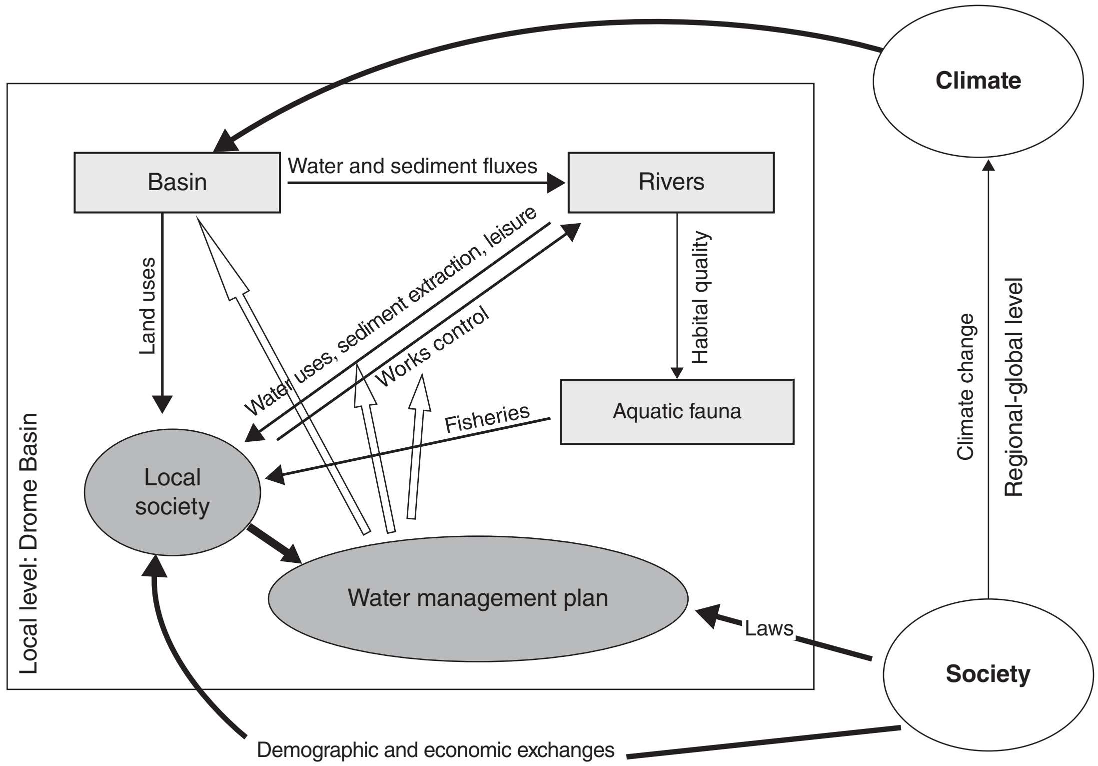


Figure 22.1 Conceptual framework of natural and anthropic factors and effects on the fluvial dynamics of the Drôme River. Source: Pont *et al*, 2009. Reproduced with permission of Springer.

on basic research, to begin applying this collective wisdom to issues of social concern.

The upsurge in the application of geomorphology has also been driven by the recognition of the costs, both financial and environmental, of ignoring natural system processes and structure in river channel management (Gilvear 1999). Legislative and economic drivers aimed at reversing a trend of ecological degradation have transformed the way in which many agencies approach intervention in river systems (Sear and Arnell 2006). Translation of science into policy frequently has long lead times and uptake of policy at the operational level is probably much longer again (Newson 1988). Furthermore, the trigger for any particular phase of uptake may be an externally imposed policy shift, which invites a subsequent scientific input, rather than an advance in science that demands policy modification. The recent policy emphasis on sustainable river channel management and 'working with natural processes' (Raven *et al.* 2002; DEFRA 2014) exemplifies a shift of stance driven by political pressure rather than scientific logic. Nevertheless, statutory requirements to have regard for 'physiographic features' or 'hydromorphology' (note the emphasis on the static descriptive nature of 'geomorphology' in legislation, which lags 30–40 years behind the shift away from this position in the discipline) and the ecological integrity of river systems have focused attention on their natural form and function. Most recently, the rise of physical habitat restoration has stimulated new research initiatives among engineers focusing on the hydraulic functions of river channel features, while ecologists are increasingly recognizing the value of geomorphology in describing and accounting for the habitat structure of aquatic systems (Jeffers 1998; Newson *et al.* 1998a; Newson and Newson 2000) and their value in delivering societal benefits – termed 'ecosystem services' (Thoms and Sheldon 2002; Thorp *et al.* 2010).

22.3 Meeting the demand: geomorphological training and application

As river managers and other scientific disciplines recognized a need for geomorphological input over the past decades, the established field of geomorphology was not prepared to meet the demand. Instead, much of this demand was met by non-geomorphologists with little academic training (at least in geomorphology) and frequently using what might be termed 'short cuts'. For example, non-geomorphologists have based channel reconstructions on relations between channel width and meander wavelength and on predictions of 'stable' channel configuration derived from a classification scheme, instead of undertaking a true geomorphological study of the river under consideration. Although these applications are often termed 'geomorphically based', they typically lack an understanding of basin-scale influences or even channel-level process interactions that actually determine the success of the intervention (Sear 1994; Sear *et al.* 2010). Moreover, they typically involve applications of only the (limited) tools to which the non-geomorphologist has been exposed.

More recently, academically trained geomorphologists have responded to the demand from managers by evolving classification and design methods using the basic research within the discipline (see Chapter 7 for a review) and by conducting post-project appraisals of restoration as a basis for improving future designs (Downs and Kondolf 2002). The challenge still remains, however, to educate a broad section of society as to the existence of the field and its real potential contributions to the management of rivers (Brookes 1995) and to communicate to practitioners alternative approaches to the 'cookbook' methods so popular now.

Restoration projects designed and implemented by professionals without a solid background in fluvial geomorphology commonly have not recognized basic but important controls on channel form, such as legacy effects of mining or flood control efforts, changed sediment supply from the catchment or even the implications of the position of the reach within the larger drainage network (e.g. depositional reaches at the transition from piedmont uplands to coastal plain along the Atlantic Seaboard of the United States). Reading the written justifications for such projects, it is clear that one of the shortcomings of the lack of substantive training is that one tends not to ask the right questions, or to use the full range of tools available. With limited training, one is likely to approach every problem in essentially the same way and one is unlikely to step back from the manager's immediate concern (be it with bank erosion or degradation of fish habitat) to redefine the problem in terms of longer term and catchment-scale processes that may be the underlying cause of the perceived reach-level problem.

22.4 The role of geomorphology in planning and management

Interactions between fluvial and human systems

Most rivers have been affected by human intervention to some extent or another, so their current conditions result from the interplay of the river and social systems (Fig. 22.2). Within the river system, flow regime (Q) and sediment load (Q_s) from the basin are the independent variables that largely determine alluvial channel form, as reflected in the adjustment of dependent variables of width, depth, grain size and pattern. This simple system can be made more complex by adding the biological and chemical elements and their relationships with the geomorphic elements. Human activities (a function of the social system) can affect both the independent variables (e.g. through urbanization and flashier runoff) and the channel form directly (e.g. by channelization or in-channel sand and gravel mining), with resulting effects on water chemistry and aquatic and riverine ecology. Because rivers are dynamic systems, such

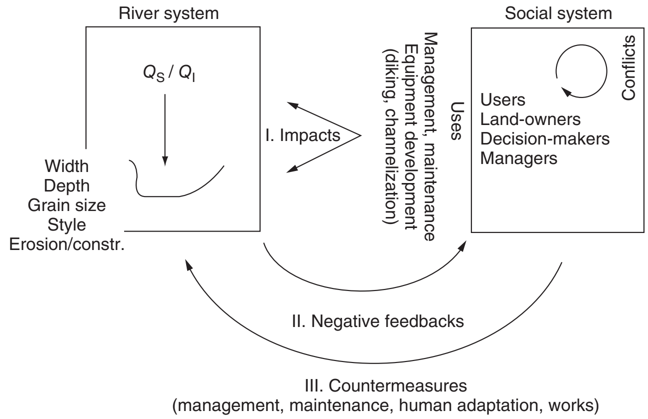


Figure 22.2 Interactions between the geomorphological river system and the social system: impacts, negative feedbacks and countermeasures.

actions typically beget reactions, such as channel incision, which in turn can affect human infrastructure or other uses (e.g. through undermining bridges and pipeline crossings) (Bravard *et al.* 1999). In response to such negative feedback from the river system, the social system tends to respond with countermeasures such as structures to control erosion of bed and banks, which in turn may produce further erosion elsewhere in the channel.

Although we speak here of the social system as a single entity, in reality, the human actors or 'stakeholders' range widely in interests, motivation and power (Kondolf and Piégay 2011). Landowners, recreational users, resource managers and elected decision-makers can act and react at different spatial and temporal scales, sometimes in complete contradiction to one another. Some conflicts recur on many rivers, such as those between canoeists and fishers, between hydroelectric companies and fish and wildlife agencies and between managers of upstream reaches and managers of downstream reaches. Social demands are complex, with multiple stakeholders and conflicts amongst them. Kondolf and Yang (2008) identified three common categories of conflicts in river restoration: among professionals, among stakeholder groups and between professionals and local groups. In this environment, fluvial geomorphologists must encourage participatory planning and management to diagnose problems and propose solutions so they can be understood by the broadest community of actors.

With river management agencies increasingly considering longer term perspective and larger spatial scales, the opportunity for geomorphologists to participate in the assessment of specific issues and to propose solutions is increasing (Piégay *et al.* 2002). Geomorphologists themselves are also social actors, influenced by their culture, history, training and experiences. They need to bear in mind that this may influence their reference system and sometimes their objectivity, so that the value system on which actions are based should be discussed and understood.

It is also worth bearing in mind that river restoration, like politics, is very much the art of the possible and restoration usually reverses only some of the anthropic changes, generally those which can be reversed without significant economic cost and/or social opposition. To facilitate floating of logs, side channels of many rivers in northern Sweden were blocked off by stone piers (Törnlund and Östlund 2002), concentrating flow in a single channel and reducing lateral connectivity. Truck transport now obviates the need for river transport and on the Pite River restoration efforts since 2001 have removed stone piers to reconnect side channels and restore lateral connectivity (Nilsson *et al.* 2005). However, a large hydroelectric dam constructed upstream in 1988 has altered the seasonal flow patterns, reducing the natural snowmelt pulse and increasing the summer–autumn baseflows, thereby artificially driving the flow regime towards a more even year-round flow, more typical of spring- or lake-fed rivers. Thus, the 'degradation trajectory' had components of both reduced lateral connectivity and reduced flow dynamics, but the restoration addressed only lateral connectivity. To restore a more natural, snowmelt flow regime would reduce hydroelectric generation. A bivariate plot shows that the degradation trajectory has vectors reflecting both decreased flow dynamics (x-axis) and lateral connectivity (y-axis), whereas the restoration trajectory has a vector along the connectivity axis only because flow regime has not been restored (Fig. 22.3). Therefore, although the full range of geomorphic, hydrological and ecological changes may be understood, societal objectives commonly limit those that a restoration programme will attempt to reverse.

Applied fluvial geomorphic questions can generally be classed as relating to (I) impacts of human development on the river system, (II) the response of the river system to these human influences or (III) the countermeasures taken by human actors to deal with the river response to development (Fig. 22.2). The success of the solutions proposed will depend in large measure on how the operators act within the social system. At level II, it is essential for them to interact with other disciplines such as ecology, economy and history to show the cascading consequences of geomorphological adjustments or functioning in term of biodiversity and recurring 'maintenance' problems, with their financial implications. At level III, scenarios must be generated to project not only the river's geomorphological response, but also resulting natural hazards, resource availability, user satisfaction and sustainable development at the basin scale. Otherwise, the solutions proposed may be effective only at a short time-scale.

How fluvial geomorphology can inform management

Applied fluvial geomorphology is now called on to evaluate the river system's function, sensitivity to change and its potential for humans. These concerns have arisen because the river is increasingly viewed as dynamic, supporting a variety of resources, and to be managed sustainably to continue providing those resources. A sampling of such questions and concerns is presented in Table 22.1. One class of questions asks, 'How does the river work?', typically posed by users who want to know if a projected action may trigger unwanted responses. Another class of questions asks, 'Why do we have such problems?' and 'Where the river is going?', typically posed to understand the causes explaining the present state and sometimes identify the actors responsible for present problems (e.g. in terms of increased flooding risk, infrastructure damage, water shortage or ecological alteration), and to evaluate the potential consequences of ongoing river adjustments to past interventions. A final class of questions, 'How can we improve the state of the river?' and 'How can we evaluate the effectiveness of policies?', encompass questions related to sustainable river restoration.

22.5 Current geomorphological practices

In current geomorphological practices, fluvial geomorphology provides management information in a set of key areas for providing decision support. We can distinguish a local (bottom-up) strategy, at the scale of the local river basin or reach concerned (the management unit), and a regional (top-down) strategy, which prioritizes actions and monitors their efficiency (such as the European Water Framework Directive) (Fig. 22.4). At a regional scale, geomorphic assessments can identify priorities and inform an adaptive strategy when previous measures were not sufficient or adapted to solve the problems.

In the bottom-up strategy, two main steps can be identified (Table 22.2):

1 Diagnosis of geomorphic state, considering both reach and basin scale perspectives and the temporal trajectory (e.g. assess causes of past evolution, present consequences and conditions, both in process-based terms).

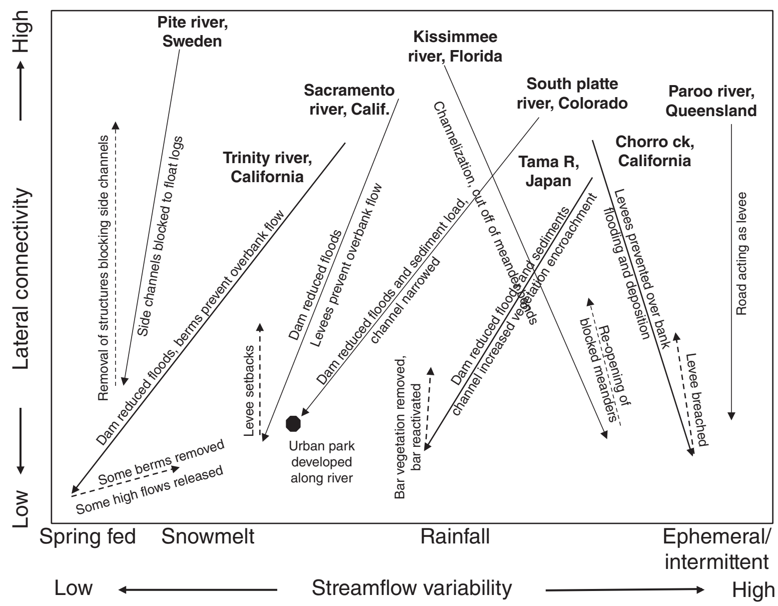
2 Geomorphic basis for project design, with three steps:

(a) project design advising on type and dimensions of channel morphology or sediment transport rate, appropriate flushing flow regimes, potential channel adjustments/responses, etc.;

(b) pre-appraisal approach allowing one to assess design efficiency, potential future changes, risk of not reaching objectives (e.g. 'sensitivity analysis');

(c) post-appraisal approach (monitoring actions), increasingly implemented to evaluate the effectiveness of public investments in restoration projects, recognizing uncertainties in channel responses and consequent success of measures.

The last few decades of investment in geomorphological research have culminated in a suite of 'standard' methods for incorporating geomorphological information into existing river management practices. These can provide a useful template for deploying the range of tools discussed in this volume (Brierley *et al.* 2002; Sear *et al.* 2010). It is essential to establish a diagnosis of the state of the river and understand the cause of the management problem and its potential consequences. The methods are designed to nest in a quasi-hierarchical fashion, collapsing from the catchment (strategic) overview of physical habitat resource, down to the project level design and assessment. This framework



(a)

(b)

Figure 22.3 (a) Diagram of trajectories of degradation and restoration for selected rivers, showing examples of change in lateral connectivity and streamflow variability. On the Pite River, plotted on the extreme left of the field, lateral connectivity was reduced by construction of stone piers to block off side channels and flow dynamics was reduced by an upstream hydroelectric dam. From Kondolf *et al.* (2006), reproduced with permission. (b). Pite River, northern Sweden, showing boulder-and-log structures built to block off side channels, to keep logs floating down the main channel. Most of these structures have been (or are being) removed to improve lateral connectivity. Photograph courtesy of Erik Tornlund, reproduced with permission.

involves the deployment of a range of geomorphological tools to provide increasing levels of certainty in the interpretation of system functioning, in support of specific management goals. The approach is based on the view of the river network as a continuum, whereby reaches are classified according to the information recovered from the catchment under study. This prevents the imposition of rigid classifications and recognizes the inherent value in the uniqueness of a river, while seeking to encourage standard approaches to the analysis of channel processes and the resulting forms and habitats.

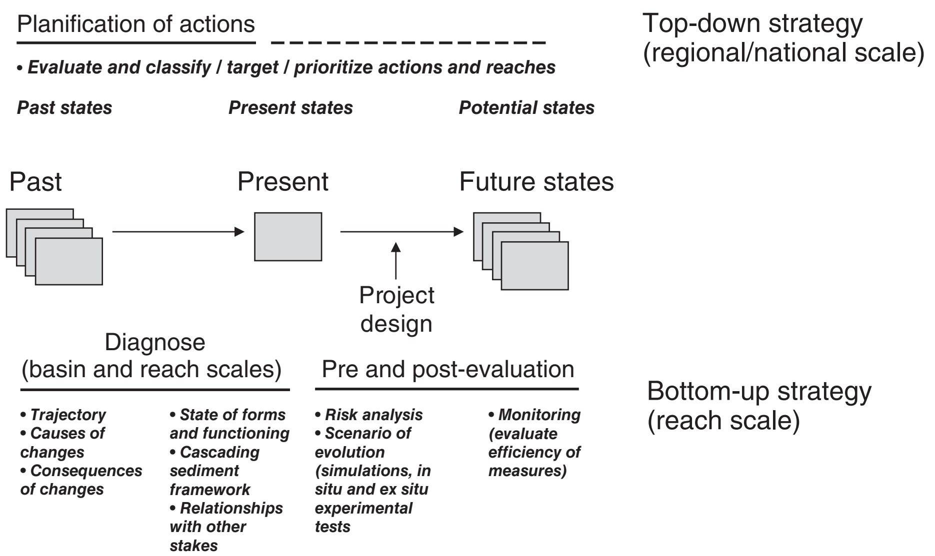
Geomorphic diagnosis

Catchment baseline surveys identify the geomorphological sensitivity and conservation status of each reach within the river network, information which is used strategically to target investment in rehabilitation or conservation designations based on

Table 22.1 Examples of geomorphological questions posed to help end-users to answer to their own questions (see Fig. 22.4).

Geomorphological questions	Reasons why geomorphological questions posed	End-users	Top-down (T) versus bottom-up approach (B)	Examples
PAST: Where the river is going? Assessment of human impacts at various spatial and temporal scales (infer potential future changes from understanding past changes)				
What is the impact of a dam on sediment transport and channel forms downstream?	Changes in fish habitat	Aquatic ecologists, fisheries management agencies	B	North Tyne: hydropower regulation impacts on spawning riffles and channel geometry: Sear (1995)
	Changes in channel geometry (narrowing, incision, aggradation)	Manager of natural hazard (flooding)	B	Hanjiang River, China: Xu (1997)
	Changes in vegetation mosaïc in the riparian zone	Landscape/aquatic ecologists and conservationists	B	Large dammed rivers in USA: Collier *et al.* (1996); Lower Ain River, France: Rollet *et al.* (2014)
Are past human actions (e.g. engineering works, mining) still inducing channel changes downstream?	Increase in channel instability	Land managers	B	Action of river maintenance activities in UK rivers: Sear *et al.* (1995)
	Geometry adjustment	River managers	B	Californian Rivers: Kondolf, (1997); English and Welsh rivers: Brookes (1987)
	Effects on biological communities	Ecologists, conservationists	B	Redwood Creek basin, NW California: Ricks (1995); Pennsylvania streams: Wohl and Carline (1996)
What is the magnitude of current and potential channel incision following channel straightening or mining?	Sensitivity of bridges to undermining	Civil engineers	B	Simon and Downs (1995)
	Drop in groundwater and impact on tree growth and dieback	Aquatic ecologists	B	Scott *et al.* (1999); Stella *et al.* (2013)
		Agriculture and water resource managers		
What is the effect of an in-channel mining site on the bedload transport and associated geomorphology of the river?	Beach degradation downstream	Land managers	B	Fiume Seccu and Figarella, Corsica: Gaillot and Piégay (1999)
	Fish habitat degradation	Engineers	B	Massive channel incision in Wooler Water: Sear and Archer (1998)
		Aquatic ecologists		
What can be the potential effects of catchment afforestation/deforestation?	Channel geometry and associated flooding risks and bank erosion	Manager of natural hazard (flooding)	B	Romero-Diaz *et al.* (2010), Piégay *et al.* (2004)
	Channel geometry and associated flooding risks and bank erosion	Manager of natural hazard (flooding; soil stability)	B	Deforestation following grazing: Liébault *et al.* (2005)
PRESENT: How does the river work? Assessment of on going processes and forms				
What is (or what will be if … ?) the sediment transport in a given reach?	Rate of reservoir filling	Water resource managers	B	Polish Carpathians: Lajczak (1996)
	Flooding frequency increase	Risk managers	B	Waiho fan, New Zealand: Davies and McSaveney (2001)
	Gravel resource availability	Gravel miners, administrators	B	Humptulips, Wynoochee and Satsop rivers of Washington state: Collins and Dunne 1(989)

What is the sensitivity of the river system to modifications of runoff and sediment load?	Assess possible consequences of development	Natural resource managers, developers	B	Moscrip and Montgomery (1997
	Sand wave and fish habitat alteration	River manager, fish ecologist	B	Yzeron basin: Navratil *et al.* (2013)
How do former river channels (e.g. oxbow lakes) vary in terms of sedimentation rate and geometry in a given reach?	Vegetation diversity and successional rates, life span of given states	Landscape ecologists, conservationists	B	Ain River corridor, France: Piégay *et al.* (2000)
Identify reaches in terms of potential channel shifting for targeting erodible corridor implementation	Target/prioritize actions	Risk managers, landscape ecologists	T	Rhône Basin: Alber (2012); Wiederkehr (2012)
What are the channel types at the regional/national scale? Assess a state to design actions	Planning	Planners	T	UK/France channel classifications: Newson *et al.* (1998b); Raven *et al.* (2002); see Chapter 7
FUTURE: How can we assess the effects of actions implemented or proposed? (e.g. analysis of future changes, mainly from present conditions)				
What are the effects of restoration practices?	Monitoring	Planners and managers	T	Lowland UK rivers: Sear *et al.* (1999)
What is the potential effect of sediment reintroduction on stream temperature and habitat diversity?	Risk analysis	Planners and managers	T	Drôme basin: Bertrand *et al.* (2013)
What are the best maintenance practice to balance flood hazard and natural quality?	Monitoring, evaluation of past actions	Managers, wildlife conservation specialists	B	Sustainable river maintenance procedure for UK rivers: Sear *et al.* (1995); French guidelines for riparian forest maintenance: Boyer *et al.* (1998); Danish streams: Iversen *et al.* (1993)
What are the risks associated with sediment reintroduction on channel stability?	Physical modelling	River managers	B	Rhine River: Die Moran *et al.* 2013
What are the risks associated with riprap removal and overerosion?	Physical and numerical modelling	River managers	B	Rhine River: Koll *et al.* 2010
How fast does sediment move downstream following reintroduction?	Field experimental monitoring and numerical modelling	River managers	B	Rhine River: Béraud (2012); Arnaud (2012)
What are the geomorphological designs to promote on a given site?	Scenario analysis (?)	Managers, conservationists	B	Mississippi streams: Shields *et al.* (1995)
What is the life span of a given restored habitat (e.g. gravel bar, former channel)?	Aquatic habitat monitoring or modelling	Managers, conservationists	B	Rhône River restoration: Riquier *et al.* (2015)



Characteristics of regional (top-down) vs local (bottom-up) strategies

	Regional (top-down)		Local (bottom-up)
Scale	Reach to network 1,000 to 50,000 km river length 5,000 to 500,000 km^2 (or greater)		Site to reach to small river basin 10 to 100 km river length 100 to 2000 km^2
Management objectives	Assess geomorphic/ecological/ chemical conditions of river reaches Target/prioritize actions	Identify/ develop actions	Diagnosis of trajectory, causes of change 'how it works' Project design Pre- and post-project appraisal
Data sources	Satellite and airborne imagery LiDAR regional surveys GIS national layers Network of stream gauges and SSC measures National-designed field campaigns (e.g. RHS)		Historical maps, photos, surveys, accounts Field measurements Hydraulic modelling In situ/ex situ experiments

Figure 22.4 General framework of geomorphic studies: diagnosis and project appraisal, top-down and bottom-up strategies.

physical habitat diversity. Output from a catchment scale survey of the River Wylye, a low gradient groundwater-dominated stream in southern England, highlighted sediment source areas and bank erosion in an easily comprehended map (Fig. 22.5). The change in land use through time is also a critical question for inferring potential geomorphic adjustments and it can be worthwhile to compare sequential aerial photographs to evaluate such changes and, where possible, make quantitative measurements yielding data sets amenable to statistical analysis. Figure 22.6 shows an analysis of land use between 1904 and 2011 in the Yzeron basin draining the western part of the city of Lyon, France. It is possible to see the progressive afforestation of the upper part of the basin and the urbanization of the lower part, both having potentially significant effects on the transfer of fine sediment in the main branch (Cottet 2005).

The geomorphic diagnosis is a field-based, reconnaissance survey, undertaken in a structured framework to provide consistency and ease of data entry and analysis within the GIS environment. Field based survey is also combined with a retrospective analysis of channel changes from historical information to establish the river trajectory. It provides an interpretation of the functioning of the river system in terms of a sediment budget and establishes links between this functioning, system morphology and the associated human stakes. The field survey includes inventories of features, coupled with assessments of materials and processes operating within the river corridor (Sear *et al.* 1995, 2010). Estimates of sediment supply and storage are calculated from measures of sediment deposits within the channel and floodplain (see Chapters 2, 9, 10, 11, 13 and 16), while supply from bank erosion is informed from measures of bank morphology and historical rates of erosion determined from historical surveys, maps and remotely sensed data (see Chapters 4 and 6). In this way, historical information is integrated with contemporary survey to establish

Table 22.2 A framework for incorporating geomorphological tools within river management projects (bottom-up strategy).

Stage	Pre-project			Project	
Procedure	*Geomorphological diagnosis* Catchment baseline study Fluvial audit		*Pre project appraisal* Present state process-based understanding	*Project design* Risk analysis and assessment of best scenarios (using models and pilot experimentations)	*Post project appraisal* Evaluate efficiency of measures/choices (monitoring)
	Assessment of historical trajectory at both basin and reach scale				
Aims	Overview of the river channel morphology and classification of geomorphological conservation value	Overview of the river basin sediment system typically aimed at addressing specific sediment-related management problems and identifying sediment source, transfer and storage reaches within the river network	To provide quantitative guidance on stream power, sediment transport and bank stability processes through a specific reach with the aim of understanding the relationships between reach dynamics and channel morphology	To design channels within the context of the basin sediment system and local processes	To assess the degree of compliance between design expectations and outcomes in terms of geomorphological processes, dimensions and morphology
Scale	Catchment (size <1–3000+ km^2)	Catchment (size <1–3000+ km^2) to channel segment	Project and adjacent reach	Project reach	Project reach
Methods	Data collation, including reconnaissance fieldwork at key points throughout catchment. Integration of data within GIS	Detailed studies of sediment sources, sinks, transport processes, floods and land-use impacts on sediment system. Historical and contemporary data sets derived from desk-based study. Integration of data within GIS	Field survey of channel form and flows; hydrological and hydraulic data, bank materials, bed sediments (GA/FA if not available), numerical modelling, field/flume/experiments	Quantitative description of channel dimensions and location of features, substrates, revetments etc. (GDA/FA/GA if not available)	Review of project aims/expectations. Compliance audit of channel against design. Re-survey of project data sets. Field survey approach
Core information	*Characterization* of river lengths on basis of morphology and sensitivity to management intervention	Identifies *range of options* and 'potentially destabilizing phenomena' (PDPs) for sediment-related river management problems	Sediment transport *rates* and morphological *stability/trends* 'Regime' approach where appropriate	The 'appropriate' *features* and their *dimensions* within a functionally-designed channel	*Extent of changes or conformity* to original project design and recommendations for mitigation options
Outputs	15–30-page report; GIS including photographs detailing conservation value and sensitivity of reaches to management actions	GIS; time chart of potentially destabilizing phenomena; report including recommendations for further geomorphological input where necessary	Quantitative guidance as to intervention (or not) and predicted impacts on reach and beyond. Identification of causes of specific problems where possible	Plans, drawings, tables and report suitable as input to quantity surveying and engineering costings. Justification for design. Explicit consideration of channel dynamics and sediment transport	Plans, tables, report. Assessment of project performance in terms of geomorphological processes and morphology/physical habitat. Recommendations for input into adaptive management
Destination	Feasibility studies for rehab/restoration	Investment/management staff, river managers or policy forums, project steering groups	River managers and project steering groups	River managers and project steering groups	River managers and project steering groups

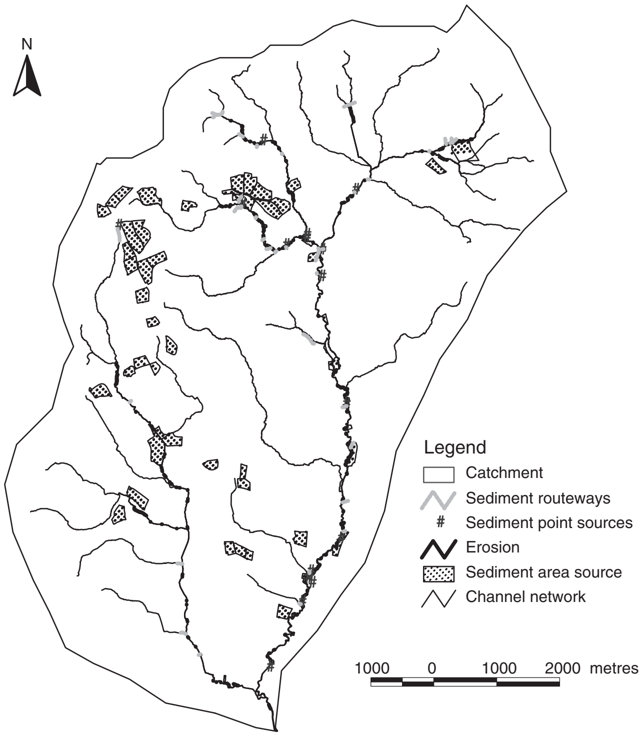


Figure 22.5 Example of a GIS output from a catchment-scale geomorphological assessment: River Wylye, a groundwater dominated river in southern England. The map highlights points of erosion, sediment point sources, more diffuses sediment source areas, and sediment routeways superimposed on the channel network.

a process-based classification based on channel activity (both vertical and lateral). Classification of the data within the GIS (see Chapters 7 and 8) facilitates the identification of zones of sediment storage, supply and transfer, informing river managers of the sources of problematic sediments or potential sediment impacts (see Chapter 5) of undertaking a given channel modification. Another layer in the information includes field determination of physical habitats and associated biological communities (Newson and Newson 2000; Sear *et al.* 2009), illuminating relationships between river geomorphology and floodplain and channel ecology, and establishes a framework for integrating different disciplines.

Geomorphic practices in project design

Geomorphological channel design uses geomorphological principles to develop an appropriate channel design. In practice, the tools deployed will depend on the nature of the design problem and the type of river system under study. For example, the restoration of a channel for physical habitat enhancement in an urban setting may be constrained in terms of what is possible in comparison with a similar scheme undertaken in relatively undeveloped landscapes. Similarly, low-energy cohesive channels may require more detailed design consideration compared with higher energy alluvial streams that are in effect able to design themselves (Sear *et al.* 2010). Approaches to geomorphological design may be based on the derivation of local hydraulic geometry relationships or from analogue reaches within the same or adjacent basins. In many situations, however, development of the catchment and modification of the hydrology and channel form may be so extensive that such approaches are not possible. In these situations, modelling of the channel form may be attempted providing that effective calibration is performed (see Chapters 17–19). Recent consideration of the process of geomorphological channel design has highlighted the role of both field survey and modelling in quantifying and reducing levels of uncertainty and communicating these to the other disciplines associated with the process.

It is useful to distinguish true geomorphological channel design from a popular approach often referred to as 'natural channel design', which involves the application of the channel classification system of Rosgen (1994) to design projects. Essentially a 'cookbook' approach to restoration, it has proved enormously popular among managers and other non-geomorphologists in the United States, being adopted by various public agencies as offering a standardized approach to prescribe restoration actions (Malakoff 2004; Lave 2008), and institutionalized as required mitigation for wetland impacts in North Carolina (Lave *et al.* 2010). In part, this popularity has derived the availability of one-week training courses where managers and staff could learn to apply the system, becoming overnight experts and ostensibly satisfying the demand for integration of geomorphology into river management without detailed geomorphic studies. Most river restoration projects designed in this way have never been objectively evaluated, but of those that have undergone post-project appraisal, the track record has included a high proportion of failures (Smith and Prestegaard 2005; Kondolf 2006) (see also Chapter 7).

Channel design requires a sound understanding of the river's functioning; hence a geomorphic diagnosis (the pre-appraisal geomorphic study) is essential as a basis for design. Once the decision has been made to implement a project, it is important to improve the process-based reach understanding and to carry out preliminary surveys before implementation.

Reach-scale, problem-focused management is often associated with specific schemes (e.g. design of river restoration projects, bank erosion control measures) and tends to involve more specific questions and requires tools to quantify system functionality (see Chapters 6 and 11–20). Thus a pre-appraisal assessment may quantify bank stability and sediment transfer within a design or 'problem' reach, while establishing it within the broader catchment context by applying a catchment-level geomorphic diagnosis. Numerous examples exist in the geomorphological literature of what could be termed 'geomorphological dynamics assessment' (e.g. Sear *et al.* 1994; Thorne *et al.* 1996). The set of tools deployed ranges from large-scale restoration programmes (where sediment load and hydrodynamics are crucial factors to quantify) with pre-project monitoring and model calibration, to small projects with modest budgets, where the tools used must be carefully selected to provide the most

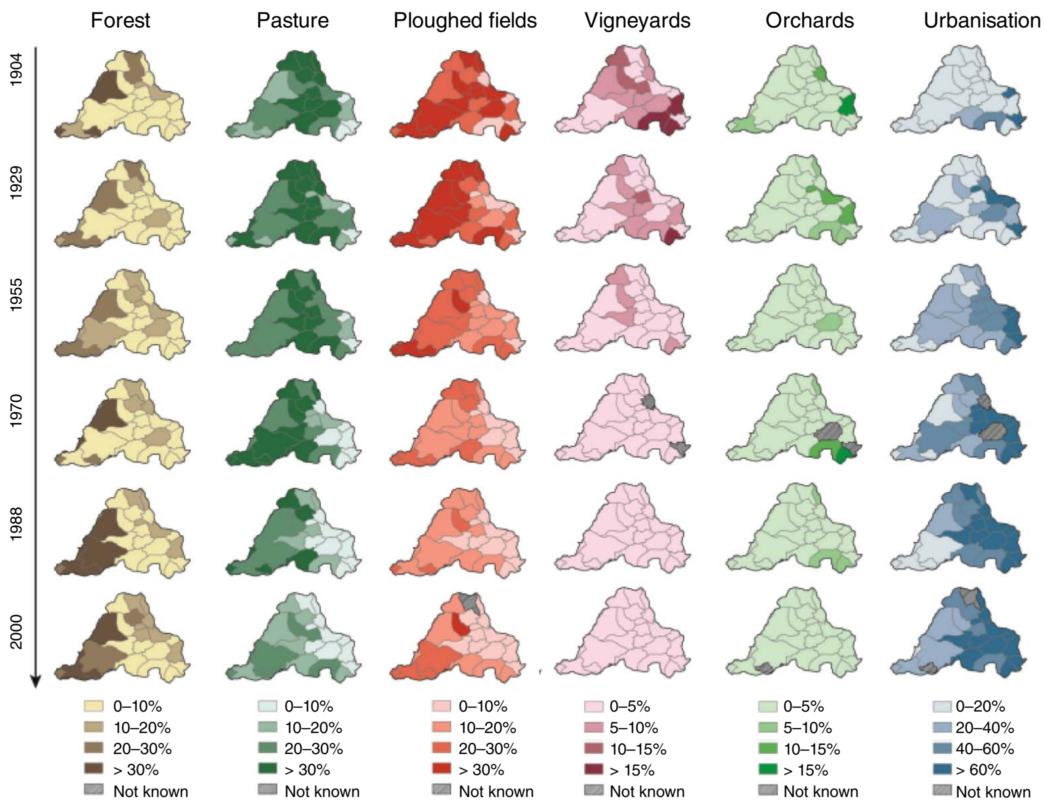


Figure 22.6 Land-use changes within the Yzeron catchment from communal statistics from 1904 to 2000. From Cottet (2005), reproduced with permission.

robust answers. A crucial point to consider here is the validity of the information obtained, particularly when legal challenge is possible, such that if more investment is necessary to answer a problem, that reality must be communicated to the stakeholders.

The *geomorphic post-project appraisal* again deploys a range of tools to determine the success of a river management programme and aims to feedback into the adaptive management process (Downs and Kondolf 2002). Such appraisal is often overlooked and under-funded by river managers, who see it as an expensive luxury rather than as a valuable tool in itself and who fear that it may reveal problems and the need to intervene in a scheme. However, even in cases where the scheme did not perform as intended, the information derived has value in terms of lessons learned that can inform future projects.

Post-project appraisal can be complex to implement, in part because of challenges in setting desired end points, which requires the definition of clear objectives and appropriate references, suitable metrics or indicators and a coherent monitoring framework (Fig. 22.7). The concept of 'reference' is still debated within the scientific community and amongst practitioners (Dufour and Piégay 2009; Morandi *et al.* 2014). References can be historically based, geographically based or process based and absolute or relative depending on whether a threshold is determined (Fig. 22.7). Historical references were probably the dominant approaches when implementing early restoration projects, the aims of which were to return to a 'pre-disturbance' state functionally and structurally (e.g. NRC 1992), with past conditions often being idealized and the environment without humans being valued. In the context of the implementation of the EU Water Framework Directive, the geographical reference is used, with the best conditions being the most natural system within a given geographical context. In restoration monitoring, a relative reference is often used. The BACI protocol (Before/After/Control/Impact) permits testing whether restoration actions have an effect independently of other factors acting at a wider scale. It is relative in the sense that it is difficult to judge whether observed changes are significant, so that thresholds must be determined as a basis for assessing success.

Models in geomorphic practices

As models are increasingly used to assess the consequences of potential actions, they have highlighted differences in objectives between managers and modellers. 'The policy or legal context [may demand] a precision in model predictions that the available knowledge cannot support', such as the requirement of water law in states of the western United States that in-stream water users claim only the minimum flow needed for a purpose, such as maintenance of channel form (Wilcock *et al.* 2003). Models can also serve to 'educate managers about the ecosystem, to identify gaps in the current knowledge … and to define plausible management scenarios that merit further evaluation' (Wilcock *et al.* 2003), as exemplified in framing the

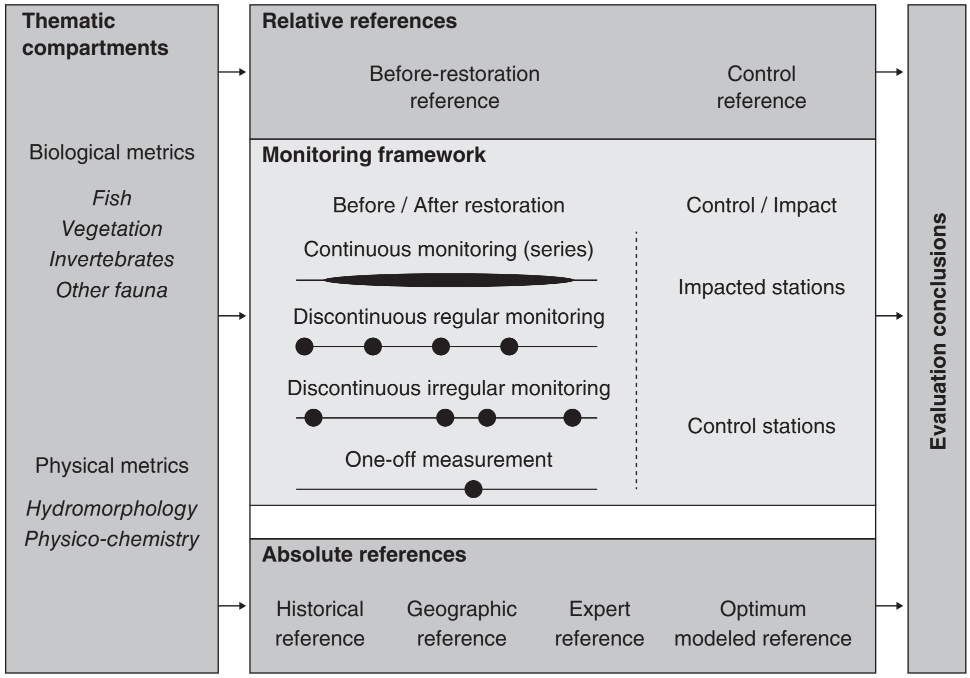


Figure 22.7 Conceptual framework of an evaluation strategy of river restoration. Morandi *et al*, 2014. Reproduced with permission of Elsevier.

range of possible dam operation alternatives along the Colorado River below Glen Canyon Dam and their potential effects on the extent of sand beaches, hydropower generation and native fish (Schmidt *et al.* 1998). The main aims are (i) to identify the best solutions, based on a good understanding of channel sensitivity to changes (e.g. a good diagnosis) and using models or experimental pilot projects to base the decision on more robust elements, (ii) to determine if it works based on a sound monitoring framework, providing the opportunity to improve or correct actions previously designed, and (iii) to target future actions on the most strategic reaches. River basin-scale planning can provide the framework with which to prioritize actions according to their urgency/degree of interest and potential system-wide benefits.

The following section takes each scale of geomorphological analysis in Fig. 22.4 and elaborates, through case studies, the application of different tools to solve specific management problems.

22.6 Case study: preventing erosion risks, from top-down to bottom-up approaches

Engineering measures to protect human structures, such as dikes, bank protection and channel straightening or deepening, affect channel geometry and bedload transport, often with negative consequences for habitat (Brookes 1988; Petts 1989). From a geomorphological point of view, bank erosion is a natural process, which contributes to the overall physical functioning of the river (Florsheim *et al.* 2008), and if stopped may result in cascading changes in channel geometry, affecting other human uses.

The 'erodible corridor' concept is to leave a wide belt within which the river channel can freely move and flood for ecological conservation and to minimize future conflicts between human settlement and bank erosion processes (Piégay *et al.* 2005). It is important to recognize that the erodible corridor approach is most effective where the river still has a dynamic flow regime and sediment load, so that it is capable of eroding, depositing, building bars, etc. The other key requirement is that there is, or can be, space for a sufficiently wide corridor: land that is either native riparian habitat or which can be returned to the river, as was the case with a strip of agricultural land along the River Aire in Geneva, Switzerland (Kondolf 2012). Generally, the more urban the site, the less room is available for an erodible corridor and the more expensive would be the land that needed to be purchased. Hence the question of what erodible corridor width should be preserved or restored – or indeed if the approach is even suitable for a given reach – involves not only geomorphological/ecological but also socioeconomic considerations (Malavoi *et al.* 1998), and can be informed by economic analysis to evaluate the relative costs of bank protection, the cost of purchasing land or easements and the annual value of any agricultural production foregone (Piégay *et al.* 1997). These factors determining the suitability of the erodible corridor approach for a given river reach can be viewed in terms to two axes: the extent to which the river has sediment and energy to move it (the y-axis in Fig. 22.8) and the space available (the x-axis in Fig. 22.8), the latter being largely a function of the extent to which the channel has been encroached by infrastructure, urban development, etc.

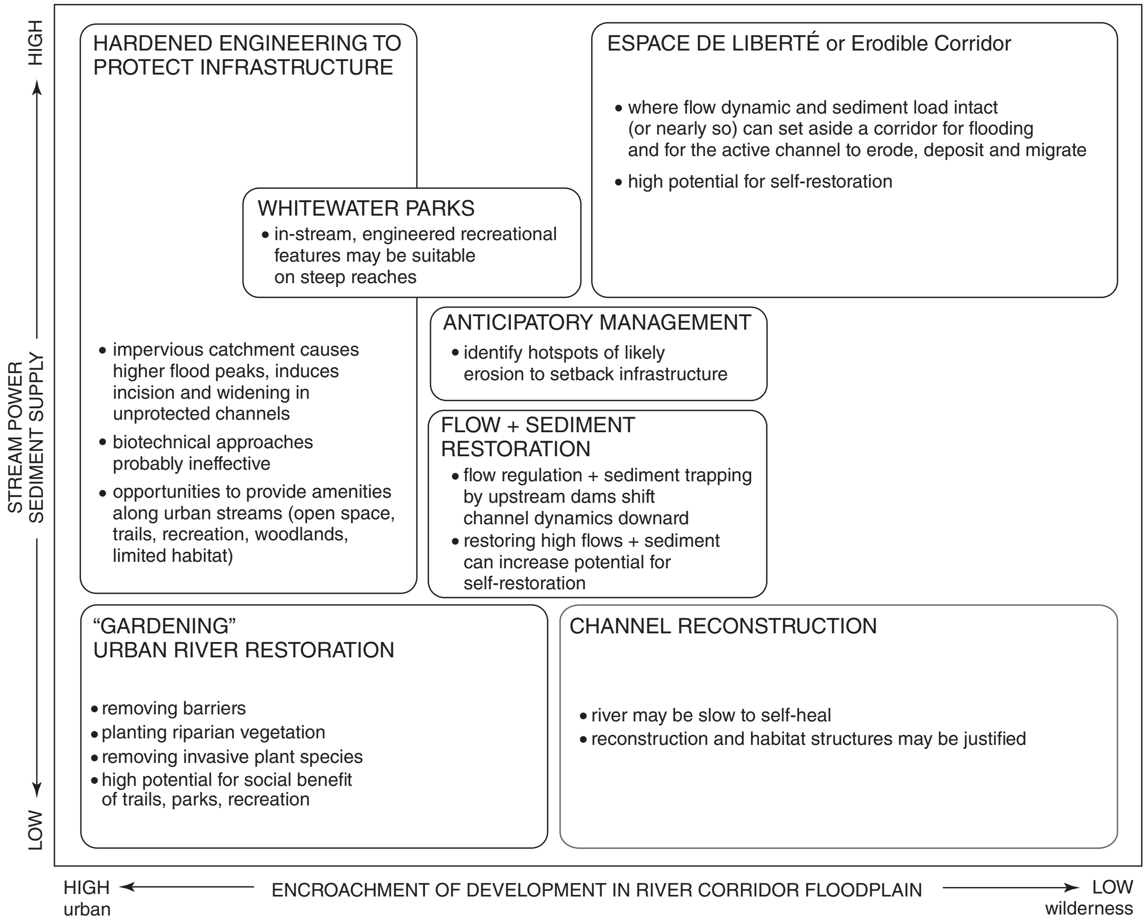


Figure 22.8 The suitability of the erodible corridor approach (or *espace de liberté*) versus other, more intrusive river management approaches, as a function of degree of urban encroachment (*x*-axis) and available stream power and sediment (*y*-axis). From Kondolf (2012), reproduced with permission.

A top-down strategy: identifying the active shifting reaches at a regional scale

In the EU, river basin authorities increasingly seek to set aside erodible corridors and thus need to identify reaches with potential for active channel shifting, which would be the best candidates for designing as erodible corridors. For the French Rhône basin, we propose a model to predict lateral erosion potential from stream power and active channel width at the scale of the 40,000 km long river network. The approach is based on a GIS procedure to map unconfined alluvial plains within which channels are potentially mobile. For a subsample of reaches, we overlay channel paths at two historical dates to establish migration rates, which we then compare with potential controlling factors (e.g. energy and sediment delivery proxies) (Fig. 22.9). This approach supports a statistical model of regional-scale channel shifting as a function of gross stream power and active channel width, rated by the catchment size. The model is then applied at an entire network scale to provide a map of channel migration potential, useful for managers in charge of targeting reaches for shifting preservation and restoration (Alber 2012; Wiederkehr 2012).

A bottom-up strategy: setting the erodible corridor

Especially on actively shifting channels (wandering, meandering and braided channels), the erodible corridor footprint can be set based on historical channel movements and projected future movements (Piégay *et al.* 2005). Historical studies of channel mobility use historical maps and aerial photographs, which are scanned, georeferenced and rectified and then analysed in a GIS platform. Using several temporal series, it is possible to overlay the different channel courses, documenting temporal and spatial variations of the channel migration rate, channel cut-off frequency and character and the areas of newly eroded and constructed floodplain, and assessing the sensitivity to erosion of individual reaches.

For example, analysis of a series of aerial photographs of the Ain River showed that from 1945 to 2000, the surface area of the unvegetated, active channel decreased from 630 to 450 ha and riparian forest established in the formerly open channel (Piégay and Saulnier 2000) (Fig. 22.10). Erosion of the floodplain surface averaged 7 ha yr^{-1} over a 40 km reach between 1980 and 1996, and 8.3 ha yr^{-1} between 1996 and 2000. Fortunately, only 6% of

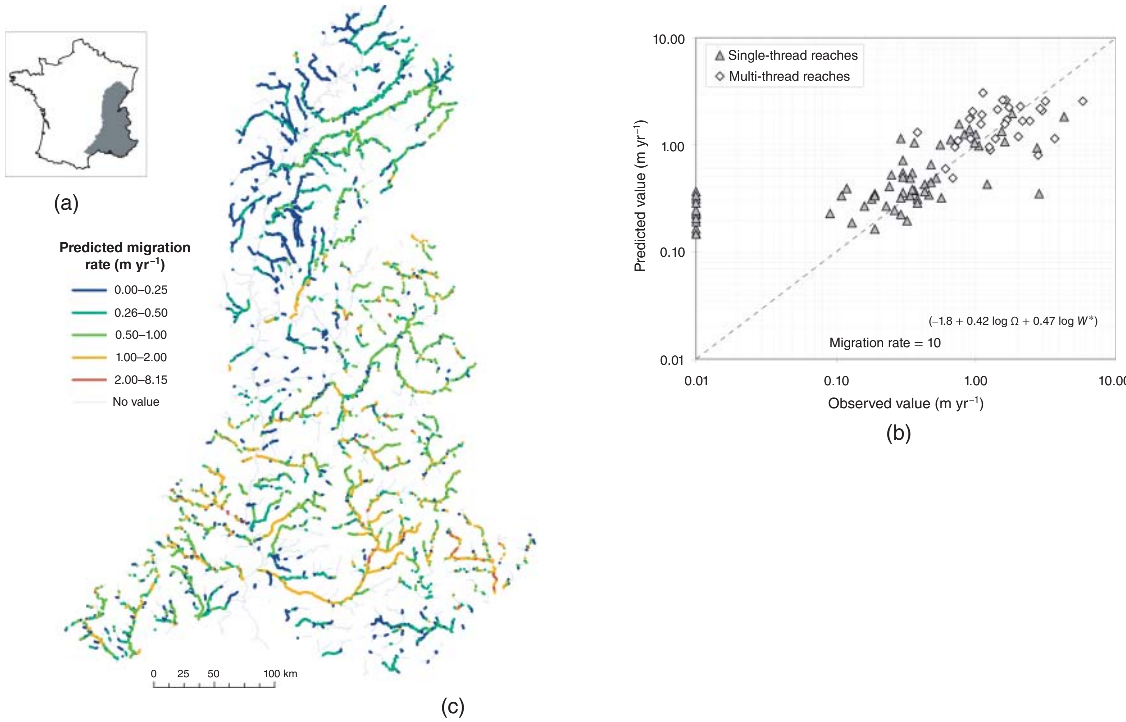


Figure 22.9 Assessment of the bank erosion susceptibility at a large basin scale for targeting erodible corridor design. (a) Location of the Rhône district in France (~90,500 km^2; 45,000 km of rivers). (b) Observed versus predicted values of mean annual bank erosion (in metres per year – log transformed) from a multiple regression with gross stream power and active channel width rated by the (catchment size)$^{0.44}$ as independent variables. Average annual bank erosion values were established on a set of about 100 reaches distributed within the catchment and on a sequence of two aerial photographs separated by 1–6 decades (average: 38 years). (c) Regional mapping of the mean annual bank erosion rate (metres per year) based on the statistical model shown in (b). See Alber and Piégay (submitted) for details. (*See plate section for color representation of this figure.*)

the eroded areas were occupied by agriculture, most of the rest consisting of the riparian forest established on the former active channel. In this context of in-channel forest establishment, low human pressure on the riparian zone and a river reach with a still dynamic flow regime and sufficient sediment supply, the erodible corridor concept (preventing development within the river corridor) has the potential to succeed. The erodible corridor zone width is based on historical analysis of channel change, with different patches distinguished according to their probability of being eroded in the next three decades. Such mapping has been done on a few tens of rivers in France, with the Ain River as one of the most advanced examples (Fig. 22.10). River managers have identified land ownership within the corridor and are developing guidelines to managing activities such as forest harvest, bridge construction or extension of existing mining sites.

Advances in numerical modelling to simulate channel evolution should improve our ability to project future channel positions, although existing tools are generally not sufficiently robust for most practical applications. Numerical modelling has progressed such that it is now possible to simulate channel meandering or braiding (see Chapter 19). Models of meander migration are based on general relationships relating lateral bank erosion rate to the near-bank velocity, channel depth and bank erodibility. Previously these models could simulate only neck cut-offs, but recent models incorporate chute cut-offs from probability functions and spatial variability of bank and floodplain erodibility. Howard's (1996) simulations of meander belt width included effects of chute cut-offs, resistant valley walls and oxbow plugs (Fig. 22.11). These models can help set erodible corridor widths and assess the residence time in the floodplain sediments (e.g. of contaminants).

22.7 Case study: pre-appraisal approach for sediment reintroduction in the Rhine: evaluating risks of restoring processes

In the Alsacian plain, the Rhine has been significantly regulated during three main phases: in the mid-19th century, the Rhine's multiple channels were converted to a single-bed, straightened and embanked channel to improve navigation; in the 1930s, the single-bed channel was narrowed by groyne fields to smooth its long profile and create consistent water levels for navigation; and finally, in the 1950s, most of the river's flow was diverted to a canal to generate electricity, with only a residual flow remaining in the natural channel.

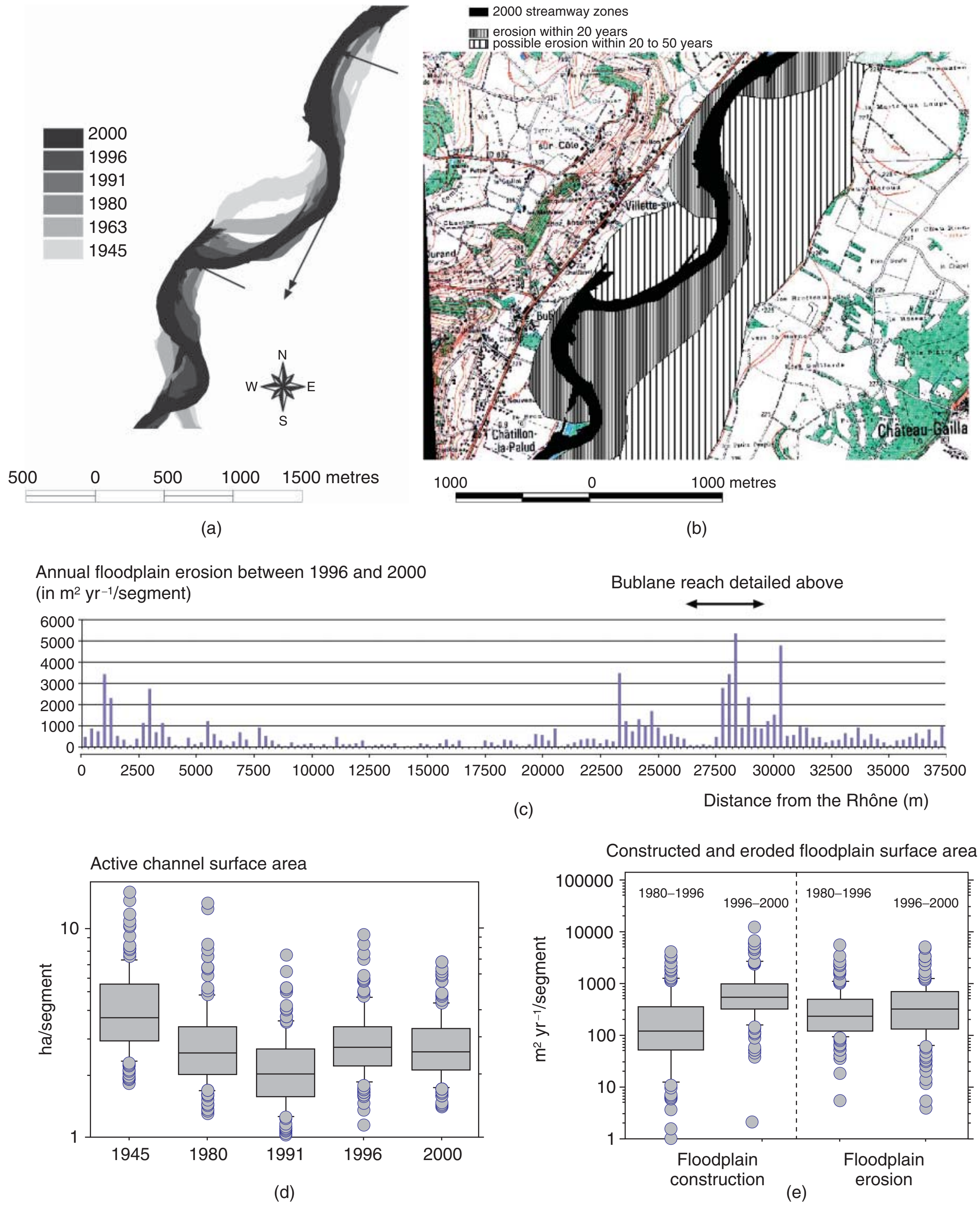


Figure 22.10 Retrospective analysis of the Ain channel mobility. View of the different channels (a), streamway differentiated according to sensitivity of zones to erosion (b), diagnosis graph showing longitudinal trends in term of recent eroded floodplain surfaces (c) and temporal trends in term of channel surface area (d) and eroded floodplain surfaces versus created floodplain surfaces (e) Segments here are river reaches of 250 m long.

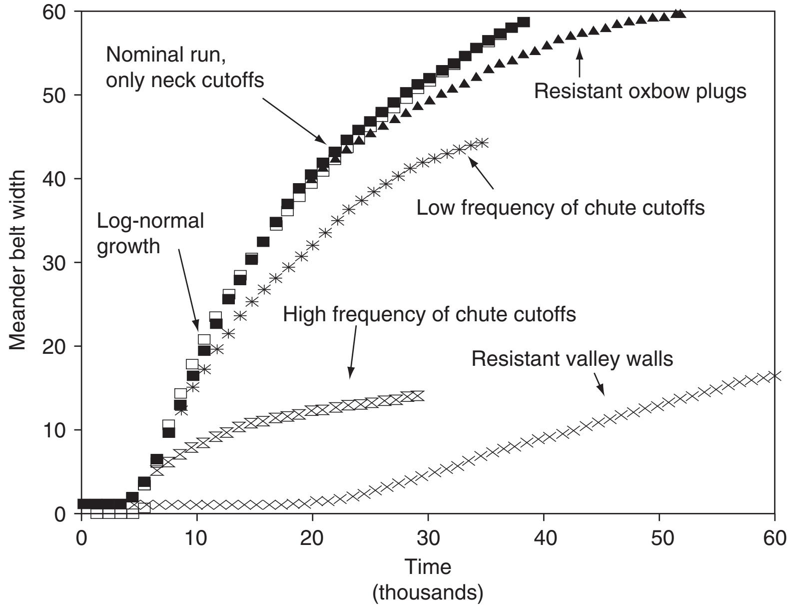


Figure 22.11 Cumulative meander belt width versus simulation time (arbitrary units). Filled boxes for simulated meandering with uniform bank erodibility and no lateral constraints. Open boxes are a logarithmic growth curve fitted to simulate results. Other curves show the effects on growth rate of chute cut-offs, resistant clay plugs and resistant valley walls (Howard 1996). Howard, 1996. Reproduced with permission of Wiley.

The reach downstream of Kembs Dam is then the longest by-passed reach (~50 km) of the Rhine. Gravel augmentation is planned here with the intent to rebuild gravel bars and other channel features for ecological benefit. A preliminary study involved a historical–geomorphological analysis (e.g. a geomorphic diagnosis) to determine the past trajectory so as to evaluate the sensitivity of the river to adjustment and a pre-appraisal approach to understand how it works currently and to assess possible negative consequences of introducing several hundred thousand cubic metres of gravel (Fig. 22.12).

This pre-appraisal approach is based on modelling and experimental field monitoring (Piégay *et al.* 2010; Arnaud *et al.* 2014, 2015). Flume experiments provided information on the potential destabilization of the bed armour and accelerated erosion of banks, while numerical modelling assessed changes in flow hydraulics and bedload transport. In addition, a field experiment involving the introduction of 22,000 m^3 of coarse sediments provided information on the rate of downstream sediment movement and development of morphological features (Fig. 22.13). Thus, the project downstream of Kembs Dam combined diagnosis and risk analysis using hydraulic modelling with both in situ and ex situ experimentation, illustrating applications of methods described in Chapters 4, 5, 6, 11, 13, 14, 15, 19 and 20. Five PhD studies were then conducted within this large European project to address different questions (Koll *et al.* 2010; Arnaud 2012; Béraud 2012; Die Moran 2012; Piquette 2014). Application of these tools permitted managers to identify the best restoration option, one that minimized risks while maximizing ecological benefits.

22.8 Case study: the River Wylye: a post-project monitoring framework to establish the performance of a range of rehabilitation schemes

Objectives

As part of a wider study of the sediment dynamics and physical habitat of the river Wylye (Fig. 22.14), a geomorphological post-project appraisal was undertaken on a range of rehabilitation schemes. Setting the performance criteria for such schemes depends on establishing their original aims. The need for rehabilitation resulted from past channelization and dredging conducted to increase floodplain drainage and to reduce frequency of overbank flooding. In many cases, the river had already been modified for milling and the retention of water levels to promote early grass growth on the floodplain. Details of the channel prior to modification were unavailable since much occurred piecemeal and as early as ~800 years BP. The result of these modifications was the removal of the gravel bed and salmon spawning habitats and the creation of channels that were over-wide and susceptible to siltation from a variety of catchment and in-channel sediment sources. Finally, channel simplification led to a reduction in habitat diversity.

The main aims of the rehabilitation were therefore to flush silt, create riffles for salmonids and to restore a physically diverse habitat. Assessing such criteria can prove problematic since few schemes have quantified targets (e.g. establish silt levels at <10% by weight of bulk sampled gravels), as was the case in this instance. Instead, an alternative approach was adopted that

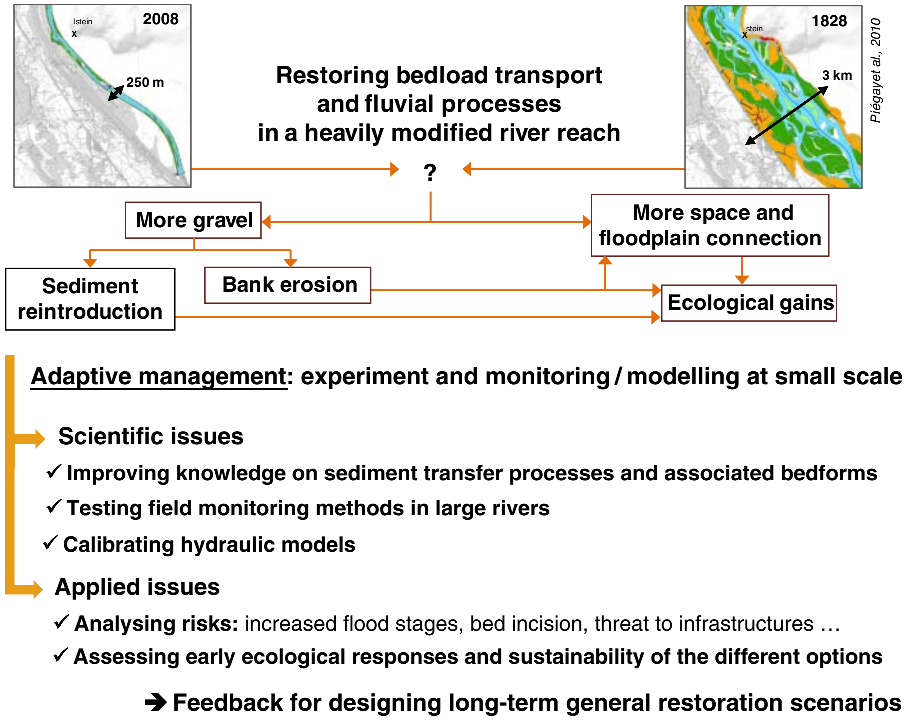


Figure 22.12 Objectives of the Rhine Redynamization Project: assess potential scenarios of restoration based on an evaluation of environmental risks, ecological benefits and sustainability of measures. Source: Piégay *et al*, 2010.

sought to establish the performance of the restored channels as measured by four criteria: (i) channel geomorphology and erosion/deposition processes, (ii) channel geometry and form, (iii) substrate heterogeneity and (iv) hydraulic habitat.

Assessment methods

The performance of three rehabilitated reaches was compared against two adjacent control reaches that had not been rehabilitated, three semi-natural reaches and one reference condition site, as a basis for assessing the overall success of each scheme. Channel geomorphology and processes were recorded through geomorphological mapping of each site, locating the features (pools, riffles, etc.) and processes (sediment storage and erosion). Channel form, channel geometry and water surface elevation (and slope) were recorded at cross-sections spaced at every bankfull channel width by standard Total Station survey. Moreover, five measurements were made at points located in the channel centre, edges and mid-way between these points, for a total of 100 points per reach. These measurements included flow depth, average velocity (measured at 0.6 depth) and substrate. Flow velocity was measured using an electromagnetic current meter that is not mechanically affected by submerged aquatic macrophytes. Substrate was estimated visually, as flow depths precluded pebble counts or photographic methods, which would have been preferred (see Chapter 13).

The geomorphological maps were used to generate indices of geomorphological and physical biotope (flow type), diversity and patchiness (sensu Newson and Newson 2000). Patchiness (the number of different features recorded) and diversity scores (estimated as the product of the number of different features and the total number of features within a reach) were normalized by reach length and compared with summary statistics and distributions of hydraulic and substrate data.

The process level analysis was based on an assessment of (i) ability to mobilize median surface bed material, indicating overall stability of the river bed, (ii) sediment continuity through the reach, indication sustainability of the reach in terms of sediment transfer and effects of rehabilitation and (iii) presence of significant bank erosion in the reach. Estimates of stream power, critical entrainment threshold for the median (D_{50}) particle motion and sediment transport rates (kg m^{-1} s^{-1}) were all established for each cross-section, in each reach for bankfull conditions using standard one-dimensional hydraulic and sediment transport modelling (see Chapter 18).

Results

At all of the sites except one semi-natural reach, bed substrates are immobile at bankfull and lower discharges, consistent with the findings of the wider 'fluvial audit', highlighting the absence of bed morphology derived from scour and deposition of coarse sediments. Rehabilitation increased sediment transport capacity and maximum mobile particle size, but not sufficiently to generate a self-sustaining coarse sediment morphology (bars, pools, riffles). Rather, at most sites sediment conveyance was limited to sediment 4 mm and finer, with sediment continuity through the reaches. At bankfull discharges, all the channels are competent

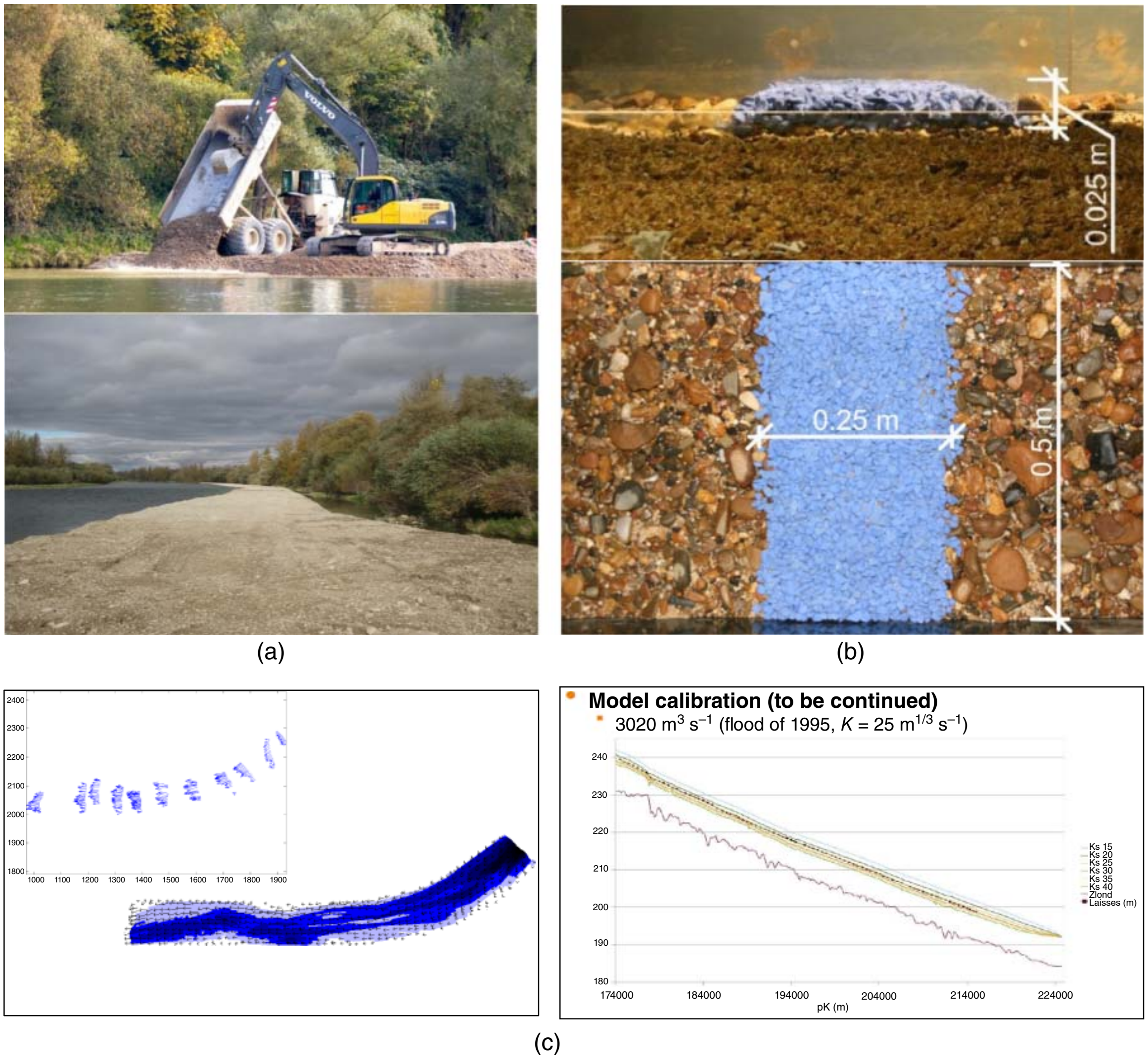


Figure 22.13 Example of tools used in a pre-appraisal geomorphic approach. (a) Field experiment with the introduction of 22,000 m^3 of gravel and its monitoring over 4 years using echo sounding, drone imagery, automatic grain sizing and particle tracking with RFID. (b) Flume experiment to reproduce armour layers and test potential effect of introduced gravel on armour layer stability. Source: A. Dittrich, Professor, Technische Universität Braunschweig. (c) 1D–2D morphodyamics model to evaluate pattern and time scale of bedload transfer downstream. Source: Jerome Le Coz, Unité de Recherche Hydrologie-Hydraulique.

to mobilize fine sediments (<2 mm), so observed accumulations of fine sediments were related to local zones of lower transport capacity, such as channel margins, backwaters, locally over-deep pools, rough vegetated channel margins and areas of flow recirculation downstream of meander bends.

In terms of channel geometry, the impacts of rehabilitation were again site specific, but were mostly reduced bankfull depth (one of the design aims) and higher and more varied width-to-depth ratios, moving the Wylye towards the typical cross-section of natural chalk streams (Sear *et al.* 1999). Most rehabilitation schemes are based on the assumption that increasing physical habitat diversity or creating a specific suite of physical habitats will increase biodiversity or specific target species. In practice, few studies have explicitly made this link. On the River Wylye, the control reaches reflecting impacts of dredging had an impoverished geomorphology relative to semi-natural reaches in the same river (Fig. 22.15). In rehabilitated reaches, the presence of coarse wood and riparian trees significantly increased the total number and type of geomorphological features present in a given length of channel, but these remained less than in semi-natural analogue reaches (Fig. 22.15). The balance of features in the rehabilitated reaches differed from that found in semi-natural chalk streams, with too few pools and berms and too many runs and woody debris limited to bankside features or 'island'-type features, not common in semi-natural chalk streams.

Overall, the rehabilitation did not significantly increase bed mobility or bank erosion. It increased fine sediment transport capacity, but at the same time increased the opportunities for accumulation due to the creation of a more varied hydraulic habitat. In terms of channel geometry, cross-section form remained simple and relatively uniform. Only where riffles were created did the long profile show significant changes, with varied hydraulic conditions and increased fine sediment loads. Rehabilitation decreased hydraulic variability whilst increasing depth variability at two sites, but each site reacted differently. At no sites did physical habitat diversity approach that found in a semi-natural stream. Future rehabilitation programmes should emphasize use of large wood to create a more varied physical habitat, but without treating the catchment-scale problems of fine sediment delivery, such rehabilitation projects will be subject to sedimentation, suggesting that a more strategic, catchment-scale approach is needed.

Figure 22.14 River Wylye geomorphological post-project appraisal. (a) A reach subjected to dredging in the 1950s for land drainage; (b) a rehabilitated reach, using soft engineering to manipulate channel form; (c) the semi-natural reference condition site.

22.9 Conclusion

The framework presented in this chapter is one of many evolving within different regions around the world. Some, such as the River Styles approach developed by Brierley *et al.* (2002), share a similar hierarchical structure in an attempt to integrate catchment-scale and reach-scale levels of investigation, whereas others are tailored to provide specific outputs for a specific purpose (see Chapter 7). What is common to all is that the application of geomorphological tools must be undertaken within a clear conceptual framework designed to identify the geomorphic principles relevant to management requirements (see Pont *et al.* 2009; Fig. 22.1). Furthermore, it is also vital in most applications to interface and transfer technology with other relevant disciplines, with GIS often serving as a useful platform. Fluvial geomorphic tools can provide information to avoid hazards (such as location of buildings in relation to bank erosion), to support uses such as navigation or resource exploitation (i.e. mining, forestry) and ecological conservation, as a foundation for long-term human benefits. Fluvial geomorphology can contribute to diagnosis, evaluation of potential impacts of proposed actions, sensitivity of systems and effectiveness of proposed measures for channel maintenance and ecological restoration.

What, then, are the challenges to better integration of fluvial geomorphology into allied fields? At this stage of the evolution of the discipline and its increasing application to solving problems, there are strong needs to articulate the benefits of the geomorphological approach, to identify indicators and metrics to monitor and assess the efficiency of measures, to learn from experience in river interventions, to develop more collaborations among geomorphological communities to benefit from experiences in different contexts and to use more of the (complementary) tools available. The development of models is a key challenge, as there are needs to simplify them, to adapt them to local context and to test their performance retrospectively. For each river, a conceptual model should be developed and the

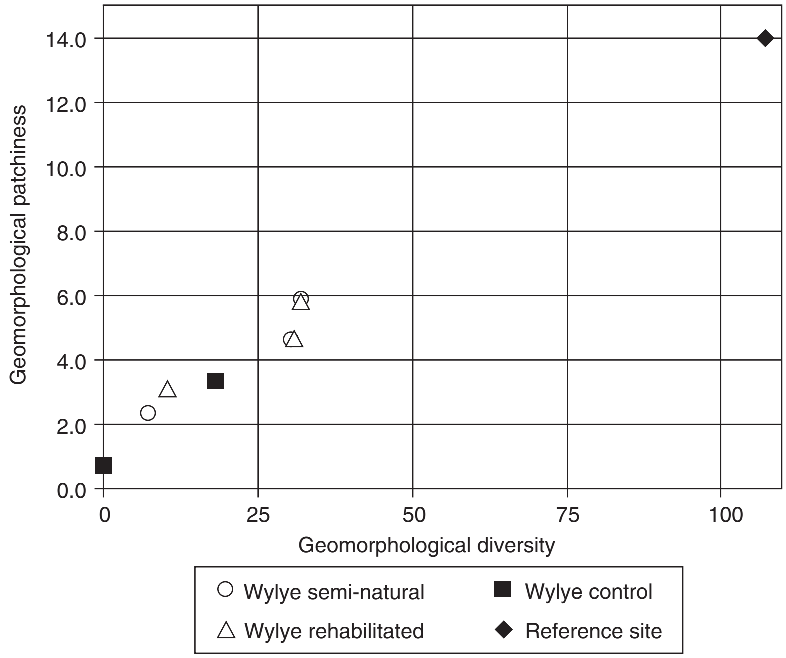


Figure 22.15 Geomorphological diversity and patchiness for control, rehabilitated and reference sites on the River Wylye. The patterns of physical habitat diversity are largely attributable to the effects of large wood.

hypothesized links tested. While geomorphologists may base actions on a clear understanding of the river's past trajectory and the understanding of the current processes and interactions, they also use a set of tools to predict future changes. Risk analysis (Bertrand *et al.* 2013), predictive modelling at increasingly large spatial and temporal scales (see the work of Lauer *et al.* 2014) and field/flume experiments are becoming essential to design sustainable improvement measures, balancing immediate and long-term goals.

Geomorphologists can provide insights into the underlying causes of 'symptoms' of ecological or stability problems in rivers, evaluate the pro and cons of different scenarios and make recommendations. However, even if the geomorphological analyses and predictions are correct, that does not guarantee a successful project because of other factors, such as cost efficiency and social acceptance. Interdisciplinary teams and scenario elaboration (prospective approaches) can help improve the chances of success of future projects.

Moreover, uncertainties will not go away and should be explicitly acknowledged and quantified to the extent possible (Sear and Darby 2008). Indeed, recent research on complex environmental systems presents evidence that the influence of multiple pressures operating at different rates can result in ecological tipping points and a failure to be able to recover back to a given state. As Sear and Arnell (2006) and others have argued, there is a strong imperative to deploy palaeohydraulic and palaeoecological tools to define the system trajectory and state relative to past states that we might seek to restore back. This has been exemplified by the work of Walter and Merritts (2012) on streams of the northeastern United States, in which their reanalysis of floodplain sediment stacks led to adoption of a different 'type' of river floodplain systems as the model for restoration. Ultimately, the success of restoration interventions depend not only on the restoration actions taken, but also on external factors such as post-project flood magnitude, ongoing river adjustments (to dams, land-use change, etc.) and counter-actions implemented elsewhere in the basin. Indeed, the importance of river adjustments to alterations in flow and sediment load is increasingly recognized. Fluvial geomorphic understanding is critically important to the understanding of likely impacts of large-scale human interventions such as downstream sediment starvation due to sediment trapping by multiple dams in a single river basin (Rubin *et al.* 2015), and this understanding in turn can motivate changes in how dams are designed and operated such that they can pass at least part of their incoming sediment load, mitigating the severity of sediment starvation downstream (Kondolf *et al.* 2014).

Although the increasing use of geomorphology is encouraging, a further problem lies in ensuring that the information is effectively translated into policy and improved practices. Information derived using tools such as those described in this book will be valuable only if the people commissioning the work understand its value and utility. Perhaps after all, among the most powerful tools available to the geomorphologist is the ability to educate non-specialists!

Land owners, flood managers, nature reserve managers, land-use planners, civil engineers, ecologists and other allied scientists can benefit from understanding geomorphological controls on basin-scale water and sediment transfer, habitat dynamics and complexity, biogeochemical cycles and water quality related to human needs. There is a clear need to explain better to end-users the geomorphic basis of management-oriented classifications and the tools that geomorphologists use for different applications.

Geomorphology programmes in universities are now training more students who can operate at a practical level and who typically work for management agencies or private companies conducting geomorphological studies and engineering designs. Moreover, interdisciplinary teams of scientists are increasingly common and the traditional boundaries between disciplines are eroding as new fields, such as ecogeomorphology, hydromorphology, sociogeomorphology, ecohydraulics and ecohydrology, develop.

Acknowledgements

Manuscript preparation was partly supported by the Beatrix Farrand Fund of the Department of Landscape Architecture and Environmental Planning, University of California, Berkeley, the EU LIFE in Rivers Project, Environment Agency and Natural England. Dr Sally German's input in the river Wylye project is gratefully acknowledged. The authors thank A. Alber, M. Cottet, J. Le Coz, A. Dittrich, B. Morandi and E. Tornlund for kind use of figures.

References

Alber, A. 2012. Etude multi-scalaire de la dynamique latérale des tronçons fluviaux. *Application au bassin rhodanien. University of Lyon, PhD Thesis*. 262 pp.

Arnaud, F. 2012. *Approches Géomorphologiques Historique et Expérimentale pour la Restauration de la Dynamique Sédimentaire d'un Tronçon Fluvial Aménagé: le Cas du Vieux Rhin Entre Kembs et Breisach (France, Allemagne)*, PhD Thesis, Université Lyon 2, France.

Arnaud, F., Piégay, H., Schmitt, L., Rollet, A.J. and Béal, D. 2014. Intérêt de la géomorphologie historique et expérimentale pour la restauration de cours d'eau: le cas du vieux Rhin en aval du barrage de Kembs (France, Allemagne). [Using historical and experimental geomorphology for restoring rivers: the case of the Old Rhine downstream of the Kembs dam (France, Germany)]. *La Houille Blanche* 4: 5–13.

Arnaud, F., Piégay, H., Schmitt, L., Rollet, A.J., Ferrier, V. and Béal, D. 2015. Historical geomorphic analysis (1932–2011) of a by-passed river reach in process-based restoration perspectives: the Old Rhine downstream of the Kembs diversion dam (France, Germany). *Geomorphology* 236: 163–177.

Béraud, C. 2012. *Modélisation Numérique des Impacts de Recharges Sédimentaires en Rivière Aménagée. Cas du Vieux-Rhin Entre Kembs et Breisach*, Thèse de Doctorat, Université Lyon 1 Claude Bernard, 248pp,

Bertrand, M., Piégay, H., Liébault, F., Pont, D. and Sauquet, E. 2013. Sensitivity analysis of environmental changes associated with riverscape evolutions following sediment reintroduction: geomatic approach on the Drôme River network, France. *International Journal of River Basin Management* 11(1): 19–32.

Boyer, M., Piégay, H., Ruffinoni, C., Citterio, A., Bourgery, C. and Caillebote, P. 1998. *Guide Technique SDAGE No. 1 – La Gestion des Boisements de Rivière: Dynamique et Fonctions de la Ripisylve*, Lyon: Agence de l'Eau Rhône Méditerranée Corse, 49pp.

Bravard, J.P., Kondolf, G.M. and Piégay, H. 1999. Environmental and societal effects of river incision and remedial strategies. In: Simon, A. and Darby, S., eds., *Incised River Channels*, Chichester: John Wiley & Sons, pp. 303–341.

Brierley, G.J., Fryirs, K., Outhet, D. and Massey, C. 2002. Application of the River Styles framework as a basis for river management in New South Wales, Australia. *Applied Geography* 22: 91–122.

Brookes, A. 1987. River channel adjustments downstream from channelization works in England and Wales. *Earth Surface Processes and Landforms* 12: 337–351.

Brookes, A. 1988. Channelized Rivers: Perspectives for Environmental Managemen*t*, Chichester: John Wiley & Sons, 326pp.

Brookes, A. 1995. Challenges and objectives for geomorphology in UK river management. *Earth Surface Processes and Lanforms* 20: 593–610.

Brookes, A. and Shields, F.D. Jr,, eds. 1996. *River Restoration: Guiding Principles for Sustainable Projects*, Chichester: John Wiley and Sons, 433pp.

Collier, M., Webb, R.H. and Schmidt, J.C. 1996. *Dams and Rivers. Primer on the Downstream Effects of Dams*, US Geological Survey Circular 1126, Tucson, AZ: US Geological Survey, 94pp.

Collins, B.D. and Dunne, T. 1989. Gravel transport, gravel harvesting and channel-bed degradation in rivers draining the southern Olympic Mountains, Washington, U.S.A. *Environmental Geology and Water Sciences* 13(3): 213–224.

Cottet, M. 2005. *Evolution de l'Occupation du Sol dans le Bassin Versant de l'Yzeron au 20éme Siècle: Premiers Éléments pour un Diagnostic Hydrologique*, Master 1 Thesis, Université Lyon 3.

Davies, T.R. and McSaveney, M.J. 2001. Anthropogenic fanhead aggradation, Waiho River, Westland, New Zealand. In: Mosley, M.P., ed., *Gravel Bed Rivers V*, Wellington: New Zealand Hydrological Society, pp. 531–553.

DEFRA (UK Department for Environment, Food and Rural Affairs). 2014. *Delivering Benefits Through Evidence. Working with Natural Processes to Reduce Flood Risk: a Research and Development Framework: Science Report*, Report SC130004/R2, Bristol: Environment Agency.

Die Moran, A. 2012. *Physical and Numerical Modelling Investigation of Induced bank Erosion as a Sediment Transport Restoration Strategy for Trained Rivers. The Case of the Old Rhin (France)*, PhD Thesis, Université Paris-Est, EDF, 243pp.

Die Moran, A., El Kadi Abderrezzak, K., Mosselman, E., Habersack, H., Lebert, F., Aelbrecht, D. and Laperrousaz, E. 2013. Physical model experiments for sediment supply to the old Rhine through induced bank erosion. *International Journal of Sediment Research* 28(4): 431–447.

Downs, P.W. and Kondolf, G.M. 2002. Post-project appraisal in adaptive management of river channel restoration. *Environmental Management* 29(4): 477–496.

Dufour, S. and Piégay, H. 2009. From the myth of a lost paradise to targeted river restoration: forget natural references and focus on human benefits. *River Research and Applications* 25: 568–581.

Florsheim, J.L., Mount, J.F. and Chin, A. 2008. Bank erosion as a desirable attribute of rivers. *Bioscience* 58(6): 519–529.

Gaillot, S. and Piégay, H. 1999. Impact of gravel-mining on stream channel and coastal sediment supply, example of the Calvi Bay in Corsica (France). *Journal of Coastal Research* 15(3):774–788.

Gilvear, D.J. 1999. Fluvial geomorphology and river engineering: future roles utilizing a fluvial hydrosystems framework. *Geomorphology* 31: 229–245.

Graf, W.L. 1996. Geomorphology and policy for restoration of impounded American rivers: what is 'natural'? In: Rhoads, B.L. and Thorn, C.E., eds., *The Scientific Nature of Geomorphology*, Chichester: John Wiley & Sons, pp. 443–473.

Gurnell, A.M., Rinaldi, M., Buijse, A.D., Brierley, G., et Piégay, H. 2016. Hydromorphological frameworks: emerging trajectories. *Aquatic Sciences*, 78(1): 135–38.

Howard, A.D. 1996. Modelling channel evolution and floodplain morphology. In: Anderson, M.G., Walling, D.E. and Bates, P.D., eds., *Floodplain Processes*, Chichester: John Wiley & Sons, pp. 15–62.

Iversen, T.M., Kronvang, B., Madsen, B.L., Markmann, P. and Nielsen, M.B. 1993. Re-establishment of Danish streams: restoration and maintenance measures. *Aquatic Conservation: Marine and Freshwater Ecosystems* 3: 73–92.

Jeffers, J.N.R. 1998. Characterisation of river habitats and prediction of habitat features using ordination techniques. *Aquatic Conservation: Marine and Freshwater Ecosystems* 8: 529–540.

Koll, K., Koll, K. and Dittrich, A. 2010. Sediment transport over static armour layers and its impact on bed stability. In: Dittrich, A., Koll, K., Aberle, J. and Geisenhainer, P., eds., *Proceedings of River Flow 2010: International Conference on Fluvial Hydraulics, Braunschweig*, Karlsruhe: Bundesanstalt für Wasserbau, pp. 929–936.

Kondolf, G.M. 1997. Hungry water: effects of dams and gravel mining on river channel. *Environmental Management* 21:533–551.

Kondolf, G.M. 2006. River restoration and meanders. *Ecology and Society* 11(2): 42 [online]: http://www.ecologyandsociety.org/vol11/iss2/art42/ (accessed 28 October 2015).

Kondolf, G.M. 2012. The espace de liberté and restoration of fluvial process: when can the river restore itself and when must we intervene? In: Boon, P. and Raven, P., eds., *River Conservation and Management*, Chichester: John Wiley & Sons, pp. 225–242.

Kondolf, G.M. and Piégay, H. 2011. Geomorphology and society. In: Gregory, K. and Goudie, A.S., eds., *The SAGE Handbook of Geomorphology*, London: SAGE Publications, pp. 105–117.

Kondolf, G.M. and Yang, C.-N. 2008. Planning river restoration projects: social and cultural dimensions. in Sear, D. and Darby, S., eds., *River Restoration: Managing the Uncertainty in Restoring Physical Habitat*, Chichester: John Wiley & Sons, pp. 43–60.

Kondolf, G.M., Boulton, A., O'Daniel, S., Poole, G., Rahel, F., Stanley, E., *et al.* 2006. Process-based ecological river restoration: visualising three-dimensional connectivity and dynamic vectors to recover lost linkages. *Ecology and Society* 11(2): 5 [online]: http://www.ecologyandsociety.org/vol11/iss2/art5/ (accessed 28 October 2015).

Kondolf, G.M., Gao, Y., Annandale, G.W., Morris, G.L., Jiang, E., Hotchkiss, R., *et al.* 2014. Sustainable sediment management in reservoirs and regulated rivers: experiences from five continents. *Earth's Future* 2(5): 256–280.

Lajczak, A. 1996. Modelling the long-term course of non-flushed reservoir sedimentation and estimating the life of dams. *Earth Surface Processes and Landforms* 21: 1091–1108.

Lauer, W., Li, C., Viparelli, E. and Piégay, H. 2014. MAST-1D: a size-specific sediment transport and tracer model with off-channel storage. Presented at the ASCE 2014 World Environmental and Water Resources Congress, 1–5 June 2014, Portland, OR.

Lave, R., 2008. *The Rosgen Wars and the Shifting Political Economy of Expertise*, PhD Thesis, University of California, Berkeley, 251pp.

Lave, R., Doyle, M. and Robertson, M. 2010. Privatizing stream restoration in the US. *Social Studies of Science* 40: 677–703.

Liébault, F, Gomez, B., Page, M. and Marden, M. 2005. Land-use change, sediment production and channel response in upland regions. *River Research and Applications* 27(7): 739–756.

Malakoff, D. 2004. The river doctor. *Science* 305: 937–939.

Malavoi, J.R., Bravard, J.P., Piégay, H., Hérouin, E. and Ramez, P. 1998. *Détermination de l'Espace de Liberté des Cours d'Eau*, Guide Technique No. 2, SDAGE RMC, 39pp.

Morandi, B., Piégay, H., Lamouroux, N. and Vaudor, L. 2014. How is success or failure in river restoration projects evaluated? Feedback from French restoration projects. *Journal of Environmental Management* 137(1): 178–188.

Moscrip, A.L. and Montgomery, D.R. 1997. Urbanization, flood frequency and salmon abundance in Puget Lowland streams. *Journal of the American Water Resources Association* 33: 1289–1297.

Navratil, O., Breil, P., Schmitt, L., Grosprêtre, L. and Albert, M.B. 2013. Hydrogeomorphic adjustments of stream channels disturbed by urban runoff (Yzeron River basin, France). *Journal of Hydrology* 485(2): 24–36.

Newson, M.D. 1988. Upland land use and land management – Policy and research aspects of the effects on water. In: Hooke, J.M., ed., *Geomorphology in Environmental Planning*, Chichester: John Wiley & Sons, pp. 19–32.

Newson, M.D. and Newson, C.L. 2000. Geomorphology, ecology and river channel habitat: mesoscale approaches to basin-scale challenges. *Progress in Physical Geography* 24: 195–217.

Newson, M.D., Clark, M.J., Sear, D.A. and Brookes, A.B. 1998a. The geomorphological basis for classifying rivers. *Aquatic Conservation: Marine and Freshwater Ecosystems* 8: 415–430.

Newson, M.D., Harper, D.M., Padmore, C.L., Kemp, J.L. and Vogel, B. 1998b. A cost-effective approach for linking habitats, flow types and species requirements. *Aquatic Conservation: Marine and Freshwater Ecosystems* 8: 431–446.

Nilsson, C., Lepori, F., Malmqvist, B., Törnlund, E., Hjerdt, N., Helfield, J.M., Palm, D., Östergren, J., Jansson, R., Brännäs, E. and Lundqvist, H. 2005. Forecasting environmental responses to restoration of rivers used as log float ways: an interdisciplinary challenge. *Ecosystems* 8: 779–800.

NRC (National Research Council). 1992. *Restoration of Aquatic Ecosystems: Science,Technology and Public Policy*, Washington, DC: NRC Committee on Restoration of Aquatic Ecosystems.

Petts, G.E. 1989. Historical analysis of fluvial hydrosystems. In: Petts, G.E., ed., *Historical Change of Large Alluvial Rivers, Western Europe, Great Britain*, Chichester: John Wiley & Sons, pp. 1–19.

Piégay, H. and Saulnier, D. 2000. The streamway, a management concept applied to the French gravel bed rivers. In: Nolan, T.J. and Thorne, C.R., eds., *CD-Rom: Gravel Bed Rivers 2000, Conference,* Christchurch, New Zealand*, 27 August–3 September*, http://geog.canterbury.ac.nz/services/carto/intro.htm (poster published).

Piégay, H., Cuaz, M., Javelle, E. and Mandier, P. 1997. A new approach to bank erosion management: the case of the Galaure river, France. *Regulated Rivers – Research and Management* 13: 433–448.

Piégay, H., Bornette, G., Citterio, A., Hérouin, E., Moulin, B. and Statiotis, C. 2000. Channel instability as control factor of silting dynamics and vegetation pattern within perifluvial aquatic zones. *Hydrological Processes* 14(16–17): 3011–3029.

Piégay, H., Dupont, P. and Faby, J.A. 2002. Questions of water resources management: feedback of the French implemented plans SAGE and SDAGE (1992–1999). *Water Policy* 4(3): 239–262.

Piégay. H., Walling, D.E., Landon, N., He, Q., Liébault, F. and Petiot, R. 2004. Contemporary changes in sediment yield in an alpine mountain basin due to afforestation (the upper Drôme in France). *Catena* 55: 183–212.

Piégay, H., Darby, S.A., Mosselmann, E. and Surian, N. 2005. The erodible corridor concept: applicability and limitations for river management. *River Research and Applications* 21: 773–789.

Piégay, H., Aelbrecht, D. and Béal, D., coords. 2010. Restauration morpho-dynamique et redynamisation de la section court-circuitée du Rhin à l'aval du barrage de Kembs (Projets INTERREG/EDF). Presented at the Congrès de la Société Hydrotechnique de France: "Environnement et Hydro-électricité", Lyon, 6–7 October 2010.

Piquette, E. 2014. *Le Rhin: "Projet Local?" Les Modalités d'Appropriation Sociale d'un Espace Fluvial Complexe*, PhD Thesis, Centre de Recherche et d'Etude en Sciences Sociales, Université de Strasbourg.

Pont, D., Piégay, H., Farinetti, A., Allain, S., Landon, N., Liébault, F., Dumont, B. and Mazet, A. 2009. Conceptual framework and interdisciplinary approach for the sustainable management of gravel-bed rivers: the case of the Drôme River basin (SE France). *Aquatic Sciences* 71(3): 356–370.

Raven, P.J., Holmes, N.T.H., Charrier, P., Dawson, F.H., Naura, M. and Boon, P.J. 2002. Towards a harmonised approach for hydromorphological assessment of rivers in Europe: a qualitative comparison of three survey methods. *Aquatic Conservation: Marine and Freshwater Ecosystems* 12(4): 405–424.

Rhoads, B.L. 1994. Fluvial geomorphology. *Progress in Physical Geography* 18(1): 103–123.

Ricks, C.L. 1995. Effects of channelization on sediment distribution and aquatic habitat at the mouth of Redwood Creek basin, Northwestern California. In: Nolan, K.M., Kelsey, H.M. and Marron, D.C., eds., *Geomorphic Processes and Aquatic Habitat in the Redwood Creek Basin, Northwestern California*, US Geological Survey Professional Paper 1454, Washington, DC: US Government Printing office, pp. Q1–Q17.

Riquier, J., Piégay, H. and Michalková, M. 2015. Hydromorphological conditions in eighteen restored floodplain channels of a large river: linking patterns to processes. *Freshwater Biology* 60: 1085–1103.

Rollet, A.J., Piégay, H., Bornette, G., Dufour, S., Lejot, J. and, Persat H. 2014. Assessment of consequences of sediment deficit on a gravel river bed downstream of dams in restoration perspectives: application of a multicriteria, hierarchical and spatially explicit diagnosis. *River Research and Applications* 30(8): 939–953.

Romero-Diaz, A., Belmonte-Serrato, F. and Ruiz-Sinoga, J.D. 2010 The geomorphic impact of afforestations on soil erosion in southeast Spain. *Land Degradation and Development* 21(2): 188–195.

Rosgen, D.L. 1994. A classification of natural rivers. *Catena* 22: 169–199.

Rubin, Z.K., Kondolf, G.M. and Carling, P. 2015. Anticipated geomorphic impacts from Mekong basin dam construction. *International Journal of River Basin Management* 13(1): 105–121.

Schmidt, J.C., Webb, R.H., Valdez, R.A., Marzolf, G.R. and Stevens, L.E.. 1998. Science and values in river restoration in the Grand Canyon. There is no restoration strategy that will improve the status of every riverine resource. *Bioscience* 48: 735–747.

Scott, M.L., Shafroth, P.B. and Auble, G.T. 1999. Responses of riparian cottonwoods to alluvial water table declines. *Environmental Management* 23: 347–358.

Sear, D.A. 1994. River restoration and geomorphology. *Aquatic Conservation: Marine and Freshwater Ecosystems* 4: 169–177.

Sear, D.A. 1995. The effects of 10 years river regulation for hydropower on the morphology and sedimentology of a gravel-bed river. *Regulated Rivers – Research and Management* 10: 247–264.

Sear, D.A. and Archer, D. 1998. The geomorphological impacts of gravel mining: case study of the Wooler Water, Northumberland UK. In: Klingeman, P., Komar, P.D. and Hey, R.D., eds., *Gravel-bed Rivers in the Environment*, Boulder, CO: Water Resources Press, pp. 325–344.

Sear, D.A. and Arnell, N.W. 2006. The application of palaeohydrology in river management. *Catena* 66(1): 164–187.

Sear, D.A. and Darby, S., eds. 2008. *River Restoration: Managing the Uncertainty in Restoring Physical Habitat*, Chichester: John Wiley & Sons, 331pp.

Sear, D.A., Darby, S.E. and Thorne, C.R. 1994. Geomorphological approach to stream stabilisation and restoration: case study of the Mimmshall Brook, Hertfordshire, UK. *Regulated Rivers* 9: 205–223.

Sear, D.A., Newson, M.D. and Brookes, A. 1995. Sediment related river maintenance: the role of fluvial geomorphology. *Earth Surface Processes and Landforms* 20: 629–647.

Sear, D.A., Armitage, P.D. and Dawson, F.D.H. 1999. Groundwater dominated rivers. *Hydrological Processes* 11(14): 255–276.

Sear, D.A., Hill, C.T, Newson, M.D., Emery, J. and Branson, J. 2009. A method for applying fluvial geomorphology in support of catchment-scale river restoration planning. *Aquatic Conservation* 9(5): 506–519.

Sear, D.A., Newson, M.D. and Thorne, C.R. 2010. *Guidebook of Applied Fluvial Geomorphology*, London: Thomas Telford, 249pp.

Shields, F.D., Knight, S.S. and Cooper, C.M. 1995. Incised stream physical habitat restoration with stone weirs. *Regulated Rivers – Research and Management* 10: 181–198.

Simon, A. and Downs, P.W. 1995. An interdisciplinary approach to evaluation of potential instability in alluvial channels. *Geomorphology* 12: 215–232.

Smith, S.M. and Prestegaard, K.L. 2005. Hydraulic performance of a morphology-based stream channel design. *Water Resources Research* 41(11): W11413.

Stella, J., Riddle, J., Piégay, H., Gagnage, M. and Tremelo, M.L. 2013. Climate and local geomorphic interactions drive patterns of riparian forest decline along a Mediterranean Basin river. *Geomorphology* 202: 101–114.

Thoms, M.C. and Sheldon, F. 2002. An ecosystem approach for determining environmental water allocations in Australian dryland river systems: the role of geomorphology. *Geomorphology* 47: 153–168.

Thorne, C.R., Allen, R.G. and Simon, A.. 1996. Geomorphological stream reconnaissance for river analysis, engineering and management. *Transactions of the Institute of British Geographers* 21: 455–468.

Thorp, J.H., Flotemersch, J.E., Delong, M.D., Casper, A.F., Thoms, M.C., Ballantyne, F., Williams, B.S., O'Neill, B.J. and Haase, C.S. 2010. Linking ecosystem services, rehabilitation and river hydrogeomorphology. *BioScience* 60: 67–74.

Törnlund, E. and Östlund, L. 2002. Floating timber in northern Sweden: the construction of floatways and transformation of rivers. *Environment and History* 8: 85–106.

Walter, R.C. and Merritts, D.J. 2008. Natural streams and the legacy of water-powered mills. *Science* 319: 299–304.

Wiederkehr, E. 2012. Apports de la géomatique pour une caractérisation physique multi-échelle des réseaux hydrographiques. Elaboration d'indicateurs appliqués au bassin du Rhône. PhD Thesis. Université de Lyon, 287 pp.

Wilcock, P.R., Schmidt, J.C., Wolman, M.G., Dietrich, W.E., Dominick, D., Doyle, M.W., *et al.* 2003. When models meet managers. *American Geophysical Union Monograph Series* 135: 27–40.

Wohl, N.E. and Carline, R.F. 1996. Relations among riparian grazing, sediment loads, macroinvertebrates and fishes in three central Pennsylvania streams. *Canadian Journal of Fisheries and Aquatic Sciences* 53: 260–266.

Xu, J. 1997. Evolution of mid-channel bars in a braided river and complex response to reservoir construction: an example from the middle Hanjiang river, China. *Earth Surface Processes and Landforms* 22: 953–965.

Index

Tools in Fluvial Geomorphology, Second Edition. Edited by G. Mathias Kondolf and Hervé Piégay.

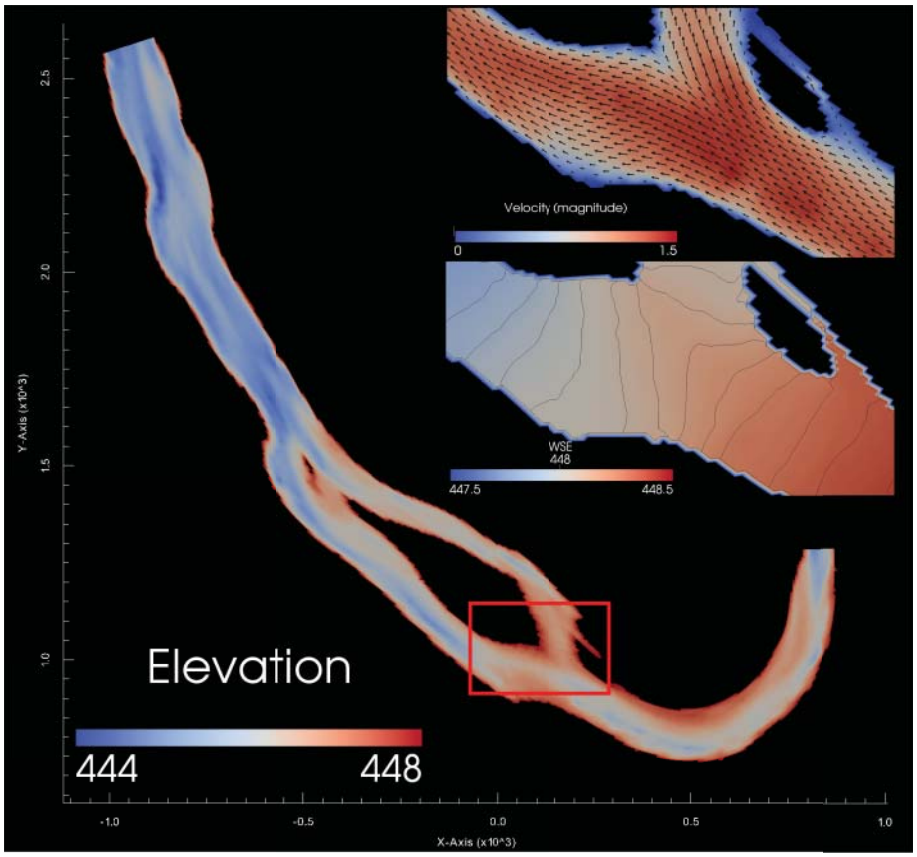


Figure 18.3 FaSTMECH model predictions for vertically averaged flow vectors (top inset) and water-surface elevation (bottom inset) on a reach of the Green River. Elevations in metres and velocity in metres per second.

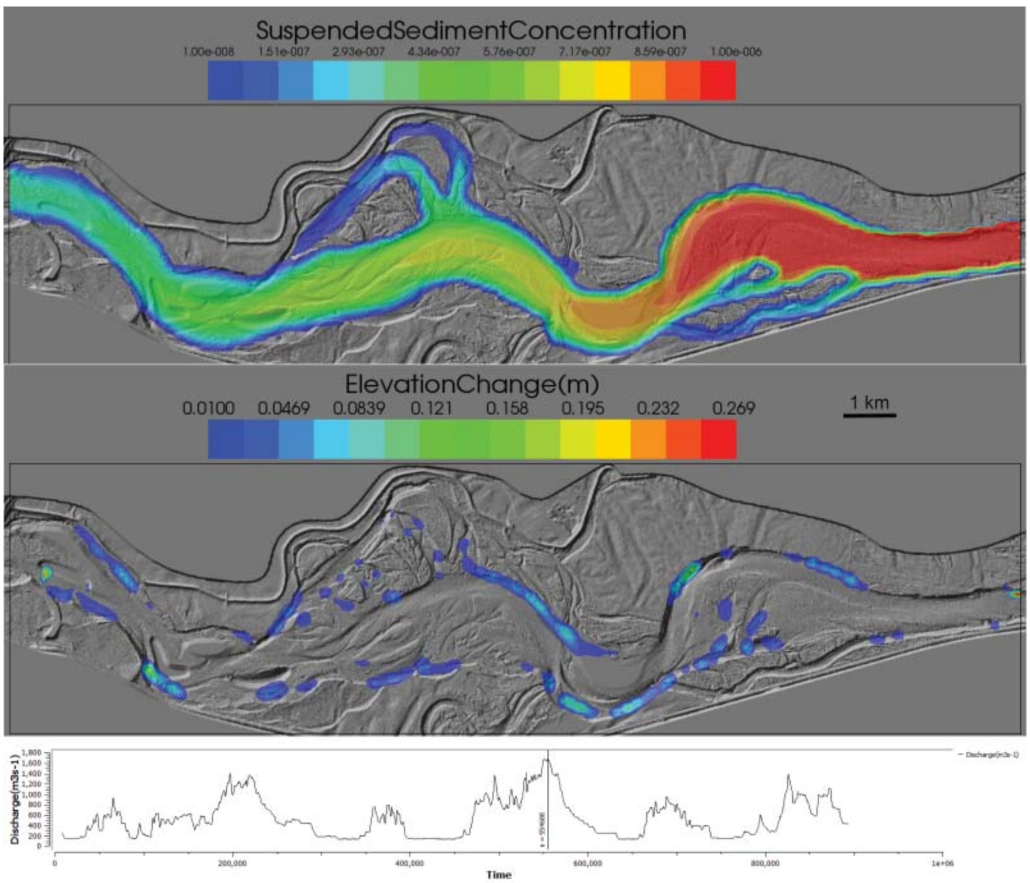


Figure 18.4 Computational prediction of the concentration and deposition thickness of fine sediment (suspended load sizes) in a reach of the Kootenai River. Time steps are 100 s, so the total time of evolution is slightly less than 3 years. The bed is shown at the time indicated by the line in the bottom diagram.

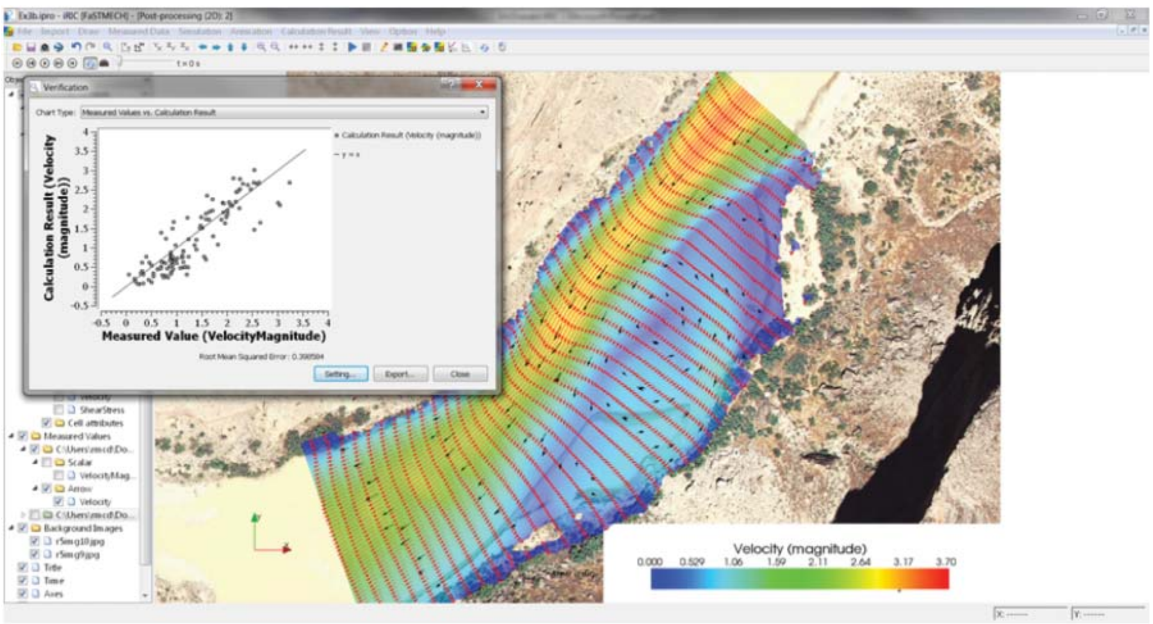


Figure 18.6 Computed vertically averaged velocity vectors compared with measured values in the Eminence Break region of the Colorado River in Grand Canyon. Calculated values are from the FaSTMECH 2.5-dimensional model in the iRIC software interface. The inset shows a comparison of the velocity magnitude between measured and calculated values. The channel is approximately 250 m wide in this view.

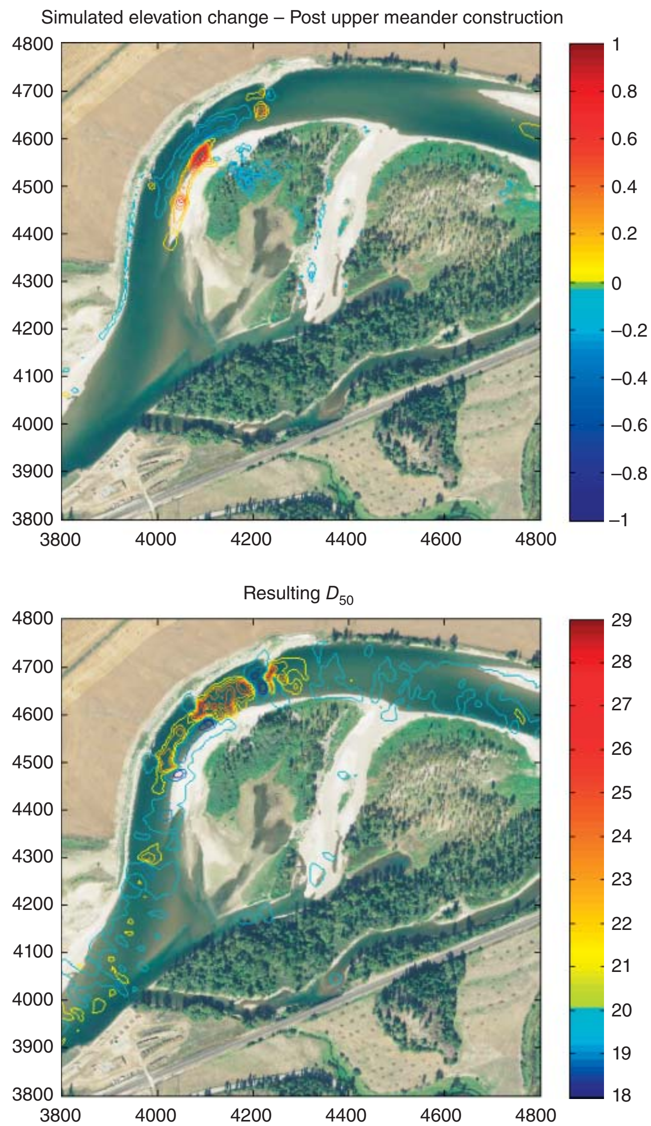


Figure 18.7 Bed morphology changes (metres) and surficial grain sizes (millimetres) predicted using FaSTMECH after the emplacement of the three spur dikes shown on the right bank.

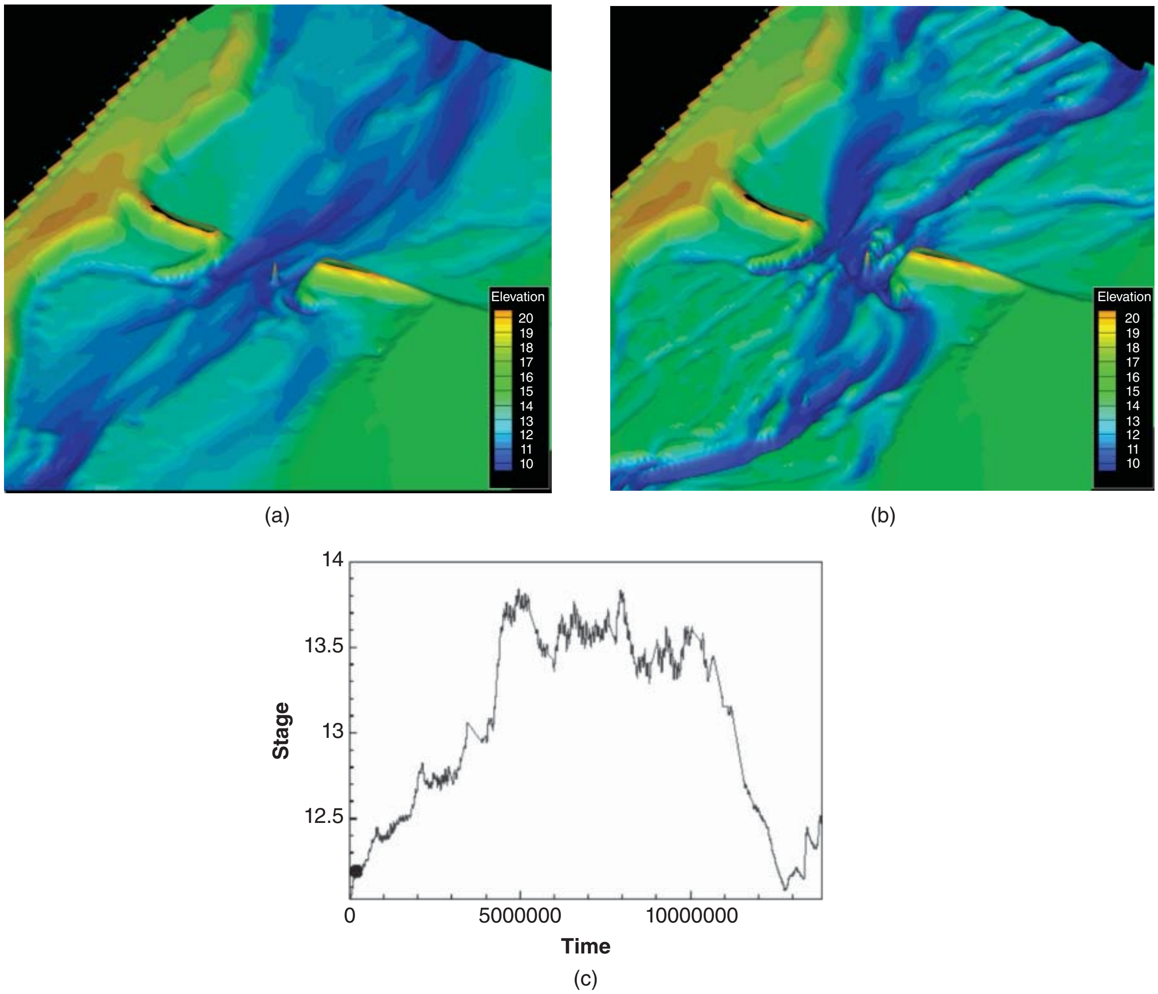


Figure 18.8 Figure showing the evolution of the Knik River channel during a high-flow event predicted using the FaSTMECH 2.5-dimensional model. (a) Pre-flood bathymetry; (b) Bathymetry after the discharge event shown in (c) (downstream flow from upper right to lower left). For scale on this perspective view, the bridge opening is ∼200 m wide. Elevations and stage in metres, time in seconds.

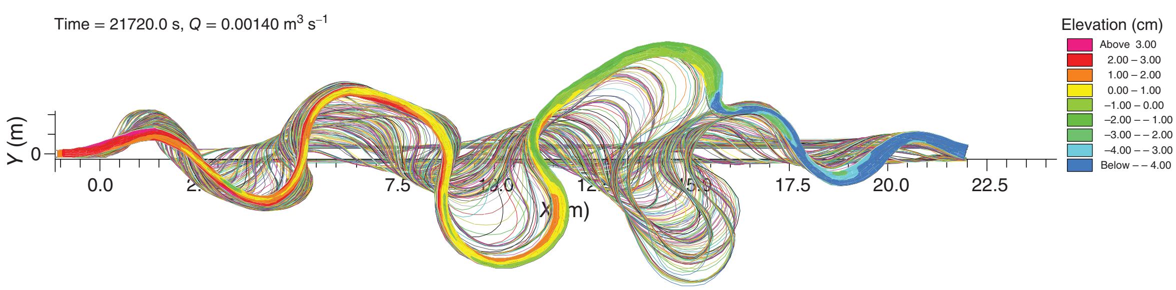


Figure 18.11 Channel planform evolution predicted starting with a straight channel with a single bend perturbation. Differently shaded lines correspond to the position of the channel at different times. Note the cutoff present near the downstream end of the reach.

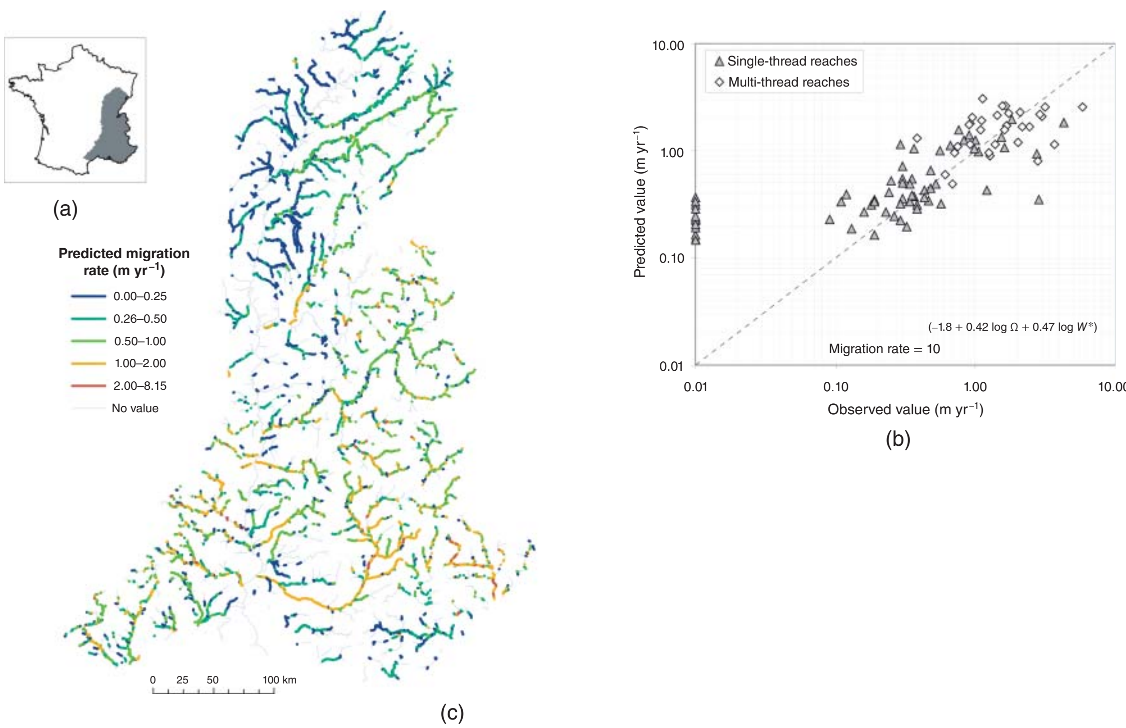


Figure 22.9 Assessment of the bank erosion susceptibility at a large basin scale for targeting erodible corridor design. (a) Location of the Rhône district in France (~90,500 km^2; 45,000 km of rivers). (b) Observed versus predicted values of mean annual bank erosion (in metres per year – log transformed) from a multiple regression with gross stream power and active channel width rated by the (catchment size)$^{0.44}$ as independent variables. Average annual bank erosion values were established on a set of about 100 reaches distributed within the catchment and on a sequence of two aerial photographs separated by 1–6 decades (average: 38 years). (c) Regional mapping of the mean annual bank erosion rate (metres per year) based on the statistical model shown in (b). See Alber and Piégay (submitted) for details.