Ellen Wohl

RIVERS IN THE LANDSCAPE

Science and Management

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Acknowledgements

Although I had contemplated writing a fluvial geomorphology text prior to starting work on this text in 2012, I put off the task as being intimidating in its complexity and requirements of time and energy. I owe the initiation of this text to the conjunction of several things. I used David Knighton's excellent *Fluvial Form and Process* as a graduate student in Vic Baker's fluvial geomorphology seminar at the University of Arizona in 1984, and when I began teaching my own graduate course in fluvial geomorphology at Colorado State University in 1989 I used this text and then the second edition published in 1998, *Fluvial Form and Process: A New Perspective*. As the volume of research being published in fluvial geomorphology steadily increased, the 1998 text became dated. Other texts with strong coverage of certain aspects of fluvial geomorphology were published, but none of them reflected the integrated approach on which the content of my graduate class increasingly relied. Consequently, the need to write a new text was nagging at the back of my mind when Martin Doyle suggested to Rachael Ballard of Wiley-Blackwell that she contact me about a new volume. I think Rachael was surprised at how readily I agreed to undertake the new volume, but the timing was right.

The organization and content of the volume were modified in response to comments from six anonymous reviewers who examined an initial extended outline. One of the reviewers commented on the proposed title, *Rivers in the Landscape*, "Where else would they be?" Indeed, but we too commonly forget the landscape context when we isolate segments of rivers for the purposes of research or management. I chose the title to emphasize the importance of the greater environmental context, from the atmosphere to the regional tectonics, from Earth history over hundreds of millions of years to human history of the past few decades, and from interactions between plants, animals, and rivers to human manipulation of rivers.

The first draft of the volume was also substantially modified in response to comments from DeAnna Laurel and in response to an extensive and thorough review by Mike Church. I thank them both for the time and effort that they put into critically reading the draft. Any errors or omissions that remain are of course mine.

In addition to those who contributed directly to this volume, I would like to acknowledge the influence of my location at Colorado State University. I am fortunate to be at an institution that puts a great deal of emphasis on water and that hosts multiple strong programs in water. Because of the ease of interacting with faculty and graduate students in civil and environmental engineering, aquatic and riparian ecology, watershed science, and geography, my approach to rivers has been transdisciplinary for many years. I would also like to acknowledge the continual nudges toward a transdisciplinary approach from my graduate students, who are likely to be more intellectually nimble in thinking about rivers outside of traditional disciplinary boundaries. Here again, my location at Colorado State is critical, as I have been able to easily accept graduate advisees with undergraduate majors as diverse as mathematics, environmental science, biology, geology, and engineering. This breadth of approach is better suited to the real breadth of river process and form, which we as scientists partition into intellectual disciplines at our own peril of artificially constraining our ability to ask and answer fundamental questions.

I would like to see all river scientists begin with the attitude expressed on a sticker on my refrigerator at home, “Yay! Rivers!,” and then follow their curiosity wherever it may lead them.

About the Companion Website

This book is accompanied by a companion website:

www.wiley.com/go/wohl/riverslandscape

The website includes:

- Powerpoints of all figures from the book for downloading
- PDFs of all tables from the book for downloading
- Supplementary materials

Chapter 1
Introduction

Rivers are the great shapers of terrestrial landscapes. Very few points on Earth above sea level do not lie within a drainage basin. Even points distant from the nearest channel are likely influenced by that channel. Tectonic uplift raises rock thousands of meters above sea level. Precipitation falling on the uplifted terrain concentrates into channels that carry sediment downward to the oceans and influence the steepness of adjacent hillslopes by governing the rate at which the whole landscape is incised. Rivers migrate laterally across lowlands, creating a complex topography of terraces, floodplain wetlands, and channels. Subtle differences in elevation, grain size, and soil moisture across this topography control the movement of groundwater and the distribution of plants and animals.

Throughout human history, people have settled disproportionately along rivers, relying on the rivers for water supply, transport, fertile agricultural soils, waste disposal, and food from aquatic and riparian organisms. People have also devoted a tremendous amount of time and energy to altering river process and form. We are not unique in this respect: ecologists refer to various organisms, from beaver to some species of riparian trees, as *ecosystem engineers* in recognition of the ability of these organisms to alter the surrounding environment. People are unique in the extent to and intensity with which we alter rivers. In many cases, river engineering has unintended consequences, and effectively mitigating these consequences requires that we understand rivers in the broadest sense, as shapers and integrators of landscape.

Geomorphologist Luna Leopold once described rivers as the gutters down which flow the ruins of continents (Leopold et al., 1964). His father, Aldo Leopold, described the functioning of an ecosystem as a round river to emphasize the cycling of nutrients and energy. Rivers can be thought of as gutters, with a strong unidirectional and linear movement of water, sediment, and other materials. Rivers can also be thought of as more broadly connected systems with bidirectional fluxes of energy and matter between the channels of the river network and the greater environment. This volume emphasizes the latter viewpoint.

1.1 Connectivity and inequality

Contemporary research and conceptual models of river form and process increasingly explicitly recognize the important of connectivity. Connectivity, sometimes referred to as coupling (Brunsden and Thornes, 1979), is multi-faceted. *Hydrologic connectivity* can refer to the movement of water down

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a hillslope in the surface and/or subsurface, from hillslopes into channels, or along a channel network (Pringle, 2001; Bracken and Croke, 2007). *River connectivity* refers to water-mediated fluxes within the channel network (Ward, 1997). *Sediment connectivity* can refer to the movement, or storage, of sediment down hillslopes, into channels, or along channel networks (Harvey, 1997; Fryirs et al., 2007; Kuo and Brierley, 2013). *Biological connectivity* refers to the ability of organisms or plant propagules to disperse between suitable habitats or between isolated populations for breeding. *Landscape connectivity* can refer to the movement of water, sediment, or other materials between individual landforms (Brierley et al., 2006). *Structural connectivity* describes the extent to which landscape units—which can range in scale from <1 m for bunchgrasses dispersed across exposed soil to the configuration of hillslopes and valley bottoms across thousands of meters—are physically linked to one another. *Functional connectivity* describes the process-specific interactions between multiple structural characteristics, such as runoff and sediment moving downslope between the bunchgrasses and exposed soil patches (Wainwright et al., 2011). Temporal variability, or connectedness of rainfall, can create spatial variability, or connectedness of flow paths, and thus functional connectivity along the slope (Wainwright et al., 2011).

Whatever form of connectivity is under discussion, the magnitude, duration, and extent of that connectivity are each important. Magnitude can be thought of as the volume of flux: is only a trickle of water moving down a channel network, or a flood? Duration describes the time span of the connectivity: can fish disperse along a river network throughout an average flow year, or only during certain seasons of high flow? Closely associated with duration is the idea of storage. If sediment stops moving downstream during periods of lower discharge, then the sediment is at least temporarily stored in the streambed and banks. Organic matter can be stored on a floodplain until overbank flows or bank erosion transport the organic matter back into the active channel. Extent is the spatial characteristic of connectivity: does sediment move readily from the crest to the toe of a hillslope, but not into the adjacent channel because the sediment is trapped and stored in alluvial fans perched on stream terraces? Recent research focuses on quantifying connectivity or developing indices of connectivity using tools such as high-resolution digital terrain models derived from aerial LiDAR (Cavalli et al., 2013) or direct measurements of fluxes (Jaeger and Olden, 2012).

These dimensions of connectivity are important for adequately characterizing fluxes within a landscape, and for understanding how human activities alter those fluxes (Kondolf et al., 2006). Many human actions substantially reduce connectivity within a river network. Dams alter hydrologic connectivity and may effectively interrupt or eliminate connectivity of sediment and some organisms along a river. Levees and bank stabilization interrupt or prevent connectivity between the channel and adjacent floodplain. Flow diversions, in contrast, may increase connectivity between drainage networks, allowing exotic organisms to migrate with the diverted water and colonize a river network. Dredging, channelization (Figure S1.1), straightening, or other activities that reduce geomorphic complexity and storage of fine sediment and nutrients typically increase longitudinal connectivity of rivers and associated downstream fluxes of sediment and solutes. By limiting overbank flows, however, these alterations reduce lateral connectivity between the channel and floodplain. Effective mitigation of undesirable human alterations of rivers requires understanding the details of connectivity.

Inextricable from connectivity is the idea of reservoirs, sinks, or storage: components of a river channel, river network, or other landscape feature in which connectivity is at least temporarily limited. Being able to quantify the magnitude and average storage time of material in flux is critical to understanding connectivity, as is being able to predict the thresholds that define upper and lower limits of storage. Sediment moving downslope from a weathered bedrock outcrop toward a stream channel might remain in storage on a debris-flow fan for 2000 years before reaching the stream channel, for example, so that the fan limits connectivity between the slope and the channel at time spans of 10^0–10^3 years. Or the sediment might progressively accumulate on the hillslope until a precipitation or seismic trigger causes the slope to cross a threshold of stability and fail in a mass movement that instantaneously introduces much of the sediment into the stream. Or the sediment might move quickly downslope and into the channel as soon as the sediment is physically detached from the bedrock outcrop, because the slope angle is too high for sediment storage. Focusing on coarse sediment transport in streams, Hooke (2003) distinguished:

- unconnected channel reaches with local sinks for sediment and lack of transport between reaches;
- partially connected reaches with sediment transfer only during large floods;
- connected reaches with coarse sediment transfer during frequent floods;
- potentially connected reaches that are competent to transfer sediment, but lack a sediment supply; and
- disconnected reaches that were formerly connected but are now obstructed by a feature such as a dam.

The point is that most natural and engineered river systems have some degree of retention of water, sediment, solutes, and organisms, and understanding net and long-term fluxes of these quantities involves quantifying both movement and storage.

Connectivity, storage, and fluxes are thus a central component of river process and form. Connectivity does not imply that all aspects of a connected valley segment, river network, or landscape are of equal importance to fluxes of matter and energy. Biogeochemists coined the phrases hot moment and hot spot. *Hot moment* describes a short period of time with disproportionately high reaction rates relative to longer intervening time periods. *Hot spot* describes a small area with disproportionately high reaction rates relative to the surroundings (McClain et al., 2003) (Figure S1.2). These ideas can be extrapolated to rivers, because any aspect of river process or form reflects inequalities in time and space.

More than 75% of the long-term sediment flux from mountain rivers in Taiwan occurs in less than 1% of the time, during typhoon-generated floods (Kao and Milliman, 2008). Approximately 50% of the suspended sediment discharged by rivers of the Western Transverse Ranges of California, USA, comes from the 10% of the basin underlain by weakly consolidated bedrock (Warrick and Mertes, 2009). Somewhere between 17% and 35% of the total particulate organic carbon flux to the world's oceans comes from high-standing islands in the southwest Pacific, which constitute only about 3% of Earth's landmass (Lyons et al., 2002). One-third of the total amount of stream energy generated by the Tapi River of India during the monsoon season is expended on the day of the peak flood (Kale and Hire, 2007). Along bedrock channels with large knickpoints, the great majority of channel incision occurs at the knickpoint.

These are but a few of the many examples that are mentioned in the remainder of this volume. Because not all moments in time or spots on a landscape are of equal importance in shaping rivers, effective understanding and management of rivers requires knowledge of how, when, and where fluxes occur.

1.2 Six degrees of connection

Any river network or segment of a single river exists in a rich and complicated context that reflects fluxes of matter and energy between the river and the greater environment, as well as the history of these fluxes. At any given moment in time, the only fluxes that are likely to be obvious are longitudinal fluxes as water and sediment move downstream. Longitudinal fluxes, however, are only one of six degrees of connection between a river and the environment (Wohl, 2010).

1. The longitudinal connection is the most obvious and intuitive (Figure 1.1; Figure S1.3a). Water, sediment, and solutes move downstream.

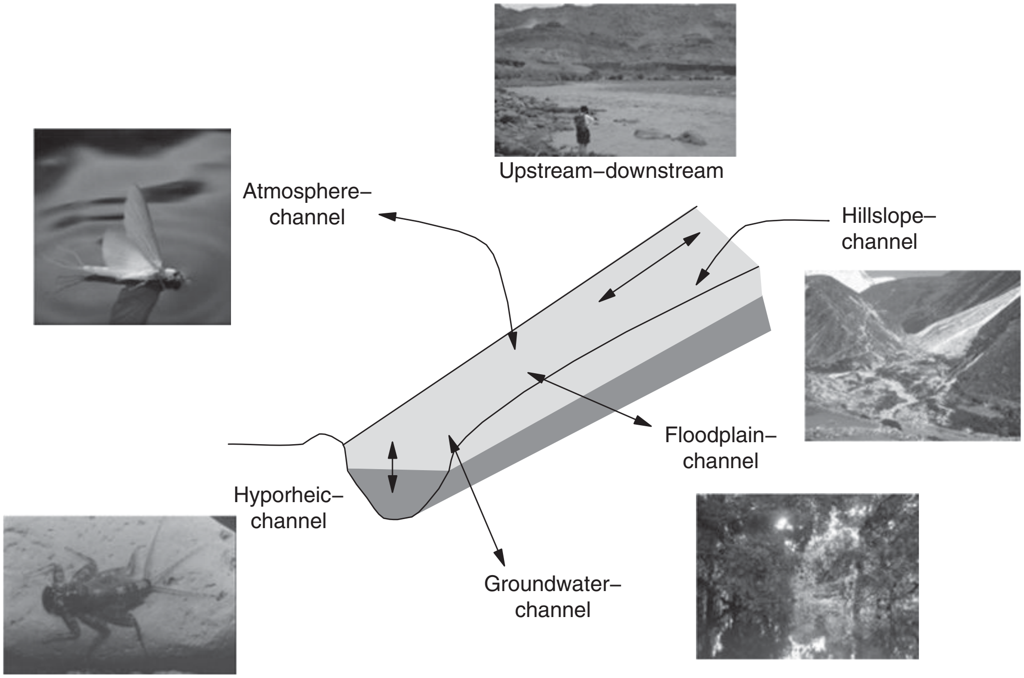

Figure 1.1 Schematic illustration of the six degrees of connection between rivers and the greater landscape. The segment of channel (lighter gray) shown here is connected to: upstream and downstream portions of the river network; adjacent uplands; the floodplain; groundwater; the hyporheic zone (darker gray); and the atmosphere. The photograph for upstream–downstream connection was taken during a flood on the Paria River, a tributary of the Colorado River that enters just downstream from Glen Canyon Dam in Arizona, USA. In this view, the Paria is turbid with suspended sediment whereas the Colorado, which is released from the base of the dam, is clear. The photograph for the hillslope–channel connection shows a large landslide entering the Dudh Kosi River in Nepal. The photograph for the floodplain–channel connection was taken along the Rio Jutai, a blackwater tributary of the Amazon River, during the annual flood in early June. In this view the "flooded forest" is submerged by several meters of water. The photograph for hyporheic–channel connection shows a larval aquatic insect (macroinvertebrate) as an example of the organisms that can move between the channel and the hyporheic environment. The photograph for atmosphere–channel connection shows a mayfly emerging from the river prior to entering the atmosphere as a winged adult (image courtesy of Jeremy Monroe, Freshwaters Illustrated).

Globally, rivers transport an estimated 7819 million tons of sediment to the oceans (Milliman and Syvitski, 1992), and approximately 0.9 Pg (1 Pg = 10^{15} g) of carbon per year (Aufdenkampe et al., 2011). Organisms move actively up- and downstream to new habitat and passively drift downstream with the current. Both European (*Anguilla anguilla*) and American eels (*Anguilla rostrata*) migrate from rivers to the Sargasso Sea off Bermuda for spawning, covering a distance of as much as 5600 km, and numerous species of salmon (*Salmo* and *Oncorhynchus* spp.) typically travel tens to hundreds of kilometers upstream from the ocean to spawn.

2. The lateral connection between the river channel and adjacent floodplain is most obvious during periods of flow with sufficient volume to overtop the banks and spread across the unchanneled valley bottom (Figure S1.3c). Water, sediment, solutes, and organisms disperse from the channel onto the floodplain during the rising and peak stages of a flood, and some of these materials concentrate once more in the channel during the falling stage of the flood.

High rates of primary production by photosynthetic organisms occur during the rising limb of the flood, providing food for the consumer organisms that follow the flood pulse onto the floodplain. High rates of decomposition occur during the flood peak, and the resulting nutrients concentrate back in the channel during the descending limb of the flood. Sediment moves onto the floodplain during the rising limb, typically remaining in storage within the floodplain until bank erosion returns the sediment to the channel (Dunne et al., 1998). Tropical river ecologists refer to the regular annual fluxes between the channel and the floodplain as the *flood pulse*, a phrase now used to refer to fluxes during floods of any recurrence interval or magnitude sufficient to create overbank flow (Junk et al., 1989; Bayley, 1991). *Flow pulses*—fluctuations in surface water below the bankfull level—create similar processes within secondary channels or areas of flow separation along a single, confined channel (Tockner et al., 2000).

Levees, channelization, and flow regulation have so restricted overbank flooding along most of the world's large and medium rivers that it is now easy to underestimate the spatial extent and duration of flooding once present along lowland rivers. The Amazon, by far the world's largest river and still one of the least engineered, can extend across 50 km of floodplain during the seasonal flood, which can last more than three months.

3. Another lateral connection is that between adjacent uplands and the river channel (Figure S1.3b). This is more likely to be a one-way flux, with water, sediment, and solutes moving downslope at the surface and subsurface into the channel. The pathways, rates and magnitudes of flux from the uplands typically exhibit substantial spatial and temporal variability. During an individual rainstorm, for example, water flowing across saturated ground may become a progressively more important source of runoff as infiltration capacity declines (Dykes and Thornes, 2000). During the dry season, soils in the seasonal tropics can develop water repellency that, along with an extensive network of macropores and pipes, facilitates rapid downslope transmission of runoff early in the wet season. Water repellency declines as the wet season continues, allowing infiltration to increase and runoff to decrease. By the peak of the wet season, however, saturated soils can promote rapid, abundant surface runoff (Niedzialek and Ogden, 2005). Rivers fed by snowmelt typically exhibit an ionic pulse when the release of solutes from the snowpack and the flushing of weathering products from the soil create the highest solute concentrations in the stream water at the initiation of snowmelt (Williams and Melack, 1991). Mineral sediment and organic matter coming from the uplands can originate in episodic, point sources such as landslides (Hilton et al., 2008a, 2008b) or via more diffuse, gradual erosion.
4. Vertical fluxes link the channel to the zone of subsurface flow immediately below the channel, with flow paths that originate and terminate at the stream (Figure S1.3e). This subsurface region is known as the *hyporheic zone*, from the Greek roots "hypo" for under or beneath and "rheo" for flow or current. Water, sediment, solutes, and small organisms such as microbes and macroinvertebrates moving between the surface and subsurface can strongly influence the volume, temperature, and chemistry of flow in the river channel, and hyporheic habitat can account for a fifth of the invertebrate production in a river ecosystem (Smock et al., 1992). The hyporheic zone can extend more than 2 km laterally from the channel in wide valleys and to depths of 10 m (Stanford and Ward, 1988).
5. Deeper vertical fluxes between the river and the saturated zone of the groundwater can also occur in both directions, with water and solutes moving into the channel in a *gaining stream* or into the groundwater in a *losing stream* (Figure S1.3d). Human activities can create gaining and losing streams. Groundwater withdrawal that lowers the water table sufficiently to prevent groundwater flow into the channel, for example, can substantially reduce stream flow in dryland rivers (Falke et al., 2011).

As in exchanges between the hyporheic zone and surface flow, exchanges between ground and surface water can influence the temperature and chemistry of river water. Solute concentrations typically increase toward saturation as groundwater moves relatively slowly through sediment or bedrock (Constantz and Stonestrom, 2003), so groundwater inputs can strongly influence river solute concentrations. The flow of rivers originating from large springs in carbonate terrains or landscapes with layered basalt flows, for example, can be almost entirely from groundwater (Gannett et al., 2003) (Figure S1.4).

Hydraulic conductivity, a measure of permeability and groundwater flow rate, can range over 12 orders of magnitude (Domenico and Schwartz, 1998). Consequently, the travel times of groundwater from areas of recharge to areas of discharge in springs or rivers can range from less than a day to more than a million years (Alley et al., 2002). This means that vertical connectivity between groundwater and channels typically influences river dynamics over long timescales relative to hyporheic flow.

6. The vertical connection between the river and the atmosphere (Figure S1.3f) can be obvious when precipitation falls directly on the river or an aquatic insect emerges from the river for the winged, terrestrial, adult phase of its life. Other fluxes involved in this connection are likely to be much less visible. Water evaporates into the atmosphere, especially from the oceans, and moves long distances before falling onto landscapes that drain into rivers. En route, the water vapor acquires very fine particulates, including: dust that may have traveled from a different hemisphere (Prospero, 1999); nitrates from vehicles, industrial emissions, and agricultural sources—the nitrates are deposited with rain and snow, and as particles and gases, in rivers hundreds of kilometers away (Heuer et al., 2000); and mercury released by vehicles and by coal-burning power plants (Grahame and Schlesinger, 2007). Volatile organic compounds—solvents such as tetrachloroethylene, chlorinated compounds such as chloroform, and others—volatilize from polluted river water into the air. Although essentially invisible, these fluxes are widespread and important.

Conceptualizing a river as having six degrees of connection with the greater environment emphasizes how diverse aspects of connectivity influence river process and form. This conceptualization also emphasizes the diversity of temporal and spatial scales across which connectivity occurs.

1.3 Rivers as integrators

Thanks to the extensive and sometimes subtle fluxes between a river and the greater environment, the forms and processes of a river integrate the physical, chemical, and biotic processes—contemporary and historical—within the environment. This may seem obvious when considering Figure 1.1, but represents the most profound summation possible regarding rivers, because of the implications.

If a river integrates diverse and seemingly unrelated processes within the greater environment, for example, then attempting to manage the river or some segment of the river in isolation from those processes is absurd.

If a river integrates … then human activities far from the physical boundaries of the channel may strongly influence the river, as when increasing atmospheric dust transport from the deserts of the southwestern United States alters snowpack melting and the resulting spring snowmelt hydrograph and water chemistry in rivers of the Rocky Mountains (Clow et al., 2002). Another example comes from the Mississippi River, where concentrations of nitrate have increased by two to five times since the early 1900s as farmers applied increasing quantities of nitrogen fertilizers to upland crop fields across the Mississippi's huge drainage basin. The resulting flux of nitrate down the river to the Gulf of Mexico tripled during the last 30 years of the twentieth century, resulting in massive algal blooms that cover a swath of the Gulf as big as New Jersey ($\sim$20,000 km^2) each year and in some years move out of the Gulf and up the eastern coast of the United States (Goolsby et al., 1999).

If a river integrates … then historical resource uses of which most people are now unaware may continue to strongly influence contemporary river process and form (Macklin and Lewin, 2008). Meandering gravel-bedded streams in the eastern United States are typically bordered by fine-grained deposits that were formerly interpreted as self-formed floodplains. Prior to European settlement, however, these river networks consisted of small anabranching channels within extensive vegetated wetlands that were buried by up to 5 m of slackwater sedimentation behind tens of thousands of seventeenth- to nineteenth-century milldams (Walter and Merritts, 2008). The ubiquitous fine sediments are thus fill terraces that reflect ongoing adjustment as the milldams breached and the channels incised. Another example comes from rivers in the Carpathian Mountains of Poland. Agriculture began in the region during the thirteenth and fourteenth centuries, and the increased sediment yield resulted in overbank aggradation along meandering rivers draining the mountains (Klimek, 1987). When the proportion of crop lands that remained bare for some portion of the year increased with more widespread cultivation of potatoes during the second half of the nineteenth century, the further increases in sediment yield caused some of the meandering rivers to assume a braided planform.

If a river integrates … then altering river process and form at one point in the river network may affect other portions of the network in unforeseen ways. The two Djerdap dams on the Danube River where it flows through Romania were built in 1970 and 1984. These massive dams, along with dozens of smaller upstream dams, have reduced sediment yields to the river's delta by 70% and silica export to the Black Sea by two-thirds relative to fluxes of these materials prior to the last third of the twentieth century. The reduced fluxes have caused erosion of the delta and a shift in the Black Sea's phytoplankton communities from siliceous diatoms to non-siliceous coccolithophores and flagellates. These changes have stimulated algal blooms and destabilized the Black Sea ecosystem (Humborg et al., 1997). Globally, humans have increased sediment supplied to and transported by rivers as a result of soil erosion, yet reduced sediment yield to the world's oceans by 1.4 billion metric tons per year because of retention behind dams (Syvitski et al., 2005). The result of this reduced coastal sediment yield has been widespread delta and near-shore erosion (Crossland et al., 2005; Yang et al., 2011).

In summary, a river integrates fluxes across a much larger and more diverse environment than the channel itself. Consequently, understanding and effectively managing river process and form is much more challenging than is likely to be recognized if a river segment is manipulated as though it was spatially and temporally isolated.

1.4 Organization of this volume

The title of this book, *Rivers in the Landscape*, reflects the inherent connections between a river and the landscape. Landscape is defined here as the physical, chemical, and biotic environment of the *critical zone*—Earth's outer layer, from the top of the vegetation canopy to the base of the soil and groundwater, that supports life. The critical zone represents the intersection of atmosphere, water, soil, and ecosystems. Recent research increasingly reminds us of what perhaps should always have been obvious: rivers do not merely flow through a landscape in isolation, but rather interact with that landscape in complex and fascinating ways. Riverine vegetation, for example, does not just increase the hydraulic resistance of overbank flow—the vegetation can alter the default river planform from braiding to meandering (Tal and Paola, 2007). Rivers do not flow passively down steep topography created by tectonic uplift—removal of mass through riverine erosion can increase the upward flux of molten rock and tectonic uplift (Zeitler et al., 2001).

Recognition of the connections between rivers and landscapes implies that the topics traditionally covered in a fluvial geomorphology text—hydraulics, sediment transport, river geometry—be treated in a manner that explicitly recognizes the influences exerted on river process and form by entities beyond the channel boundaries. Consequently, this book builds from traditional understanding of rivers toward the larger, more comprehensive viewpoint.

Chapter 2 covers the development of channels and channel networks, including how water, sediment, and solutes are produced; how they move from uplands into channels; how channel heads form; and how channel networks extend across the landscape. This chapter addresses the processes by which water moves across and through unchannelized hillslopes and concentrates sufficiently to create channels.

Chapter 3 covers channel processes, with a focus on energy (hydraulics) and quantities (hydrology). Knowledge of the basic mechanics of channelized flow is integral to understanding sediment erosion, transport and deposition, and adjustment of channel form.

Chapter 4 covers the movement of sediment in channels. The discussion begins with the sediment texture of channel beds and the processes that initiate motion of non-cohesive and cohesive sediment. Once sediment is mobilized from the streambed and banks, it can be transported in solution, in suspension, or in contact with the bed, and can be organized into bedforms.

Chapter 5 addresses channel form, exploring how water and sediment movement shape channel geometry through time and space. Interactions between process and form are implicit throughout Chapters 3 and 4, but Chapter 5 explicitly examines feedbacks between process and form at increasingly larger spatial scales, from cross-sectional geometry, through channel planform, and longitudinal gradient, to downstream trends along a river and across a river basin.

Chapter 6 summarizes process and form of fluvially created and maintained features outside of the immediate channel—floodplains, terraces, alluvial fans, deltas, and estuaries. These river landforms both reflect and influence channel process and form.

Humans and human influences on the landscape are now ubiquitous. Although human influences on rivers are mentioned where appropriate in each of the first six chapters, these influences are the explicit focus of Chapter 7, which summarizes the depth and breadth of human alterations of rivers.

Chapter 8 metaphorically steps back to use the knowledge of process and form developed in preceding chapters as a means to understand rivers in a landscape context. This chapter starts with a discussion of how topography influences the spatial distribution of river networks and energy expenditure within rivers, how rivers influence rates of landscape denudation, and the indicators used to infer relations between rivers and landscape evolution. Spatial differentiation of geomorphic process and form within river basins is discussed, followed by a re-examination of connectivity. Distinctive river characteristics associated with high and low latitudes and arid regions provide examples of the importance of landscape context, as do a series of case studies. These case studies illustrate how place-specific details of geology, climate, and land use history influence river process and form, as well as the management implications of these influences.

One of the challenges in writing a reasonably concise fluvial geomorphology text is the tremendous volume of research conducted on rivers within the past century. Scientists from diverse backgrounds in geology, geography, civil engineering, and other disciplines study river process and form via

- direct measurements and experimental manipulations of real rivers;
- indirect measurements using remote sensing imagery from space-based (e.g., aerial photographs, satellite imagery, airborne LiDAR) and ground platforms (e.g., ground penetrating radar);
- physical experiments in a laboratory;
- numerical models; and
- integration of all of these approaches.

Another fundamental challenge is the diversity of rivers. Water flows downslope under the influence of gravity. The basic physics are the same in any environment, but the ability to generalize beyond the most basic level is typically obscured by the local, place-specific details and history of a particular river. As fluvial geomorphology continues to develop as a discipline, there remains an underlying tension within the community between investigators who emphasize quantification as a means of identifying physical principles and mechanisms acting across a range of specific landscapes (Dietrich et al., 2003) and investigators who emphasize the use of historical and sedimentary records as a means of identifying the role of contingency and site-specific characteristics in river process and form.

Until perhaps the 1960s or 1970s, the great majority of river research focused on medium-sized, low- to medium-gradient, sand-bed rivers. These were the most accessible rivers for scientists living primarily in the temperate latitudes, and the foundational research conducted on these rivers gave rise to widely used conceptual models and equations for hydraulics, sediment transport, and channel geometry. As investigators have subsequently spent more time quantitatively examining rivers with steeper gradients and more resistant boundaries (gravel-bed rivers, bedrock rivers, mountain rivers) and greater hydrologic variability (seasonal tropics, drylands), as well as rivers at higher (boreal, arctic) and lower (tropical) latitudes, the ability of the "classic" models and equations to adequately describe process and form across the known spectrum of rivers becomes weaker. Throughout this volume, I explicitly address some of the unique characteristics of rivers beyond temperate zone sand-bed channels.

My intent in this text is to maintain conciseness while reflecting the diversity of natural rivers and the methods of studying rivers. More extensive discussions of this diversity are provided in supplemental material on the accompanying website. Along these lines, the references cited in the text are not an exhaustive list, but rather a starting point that combines some foundational studies and particularly integrative or insightful recent studies. The website provides a more thorough list of references for many of the topics covered in the text.

1.5 Understanding rivers

Recent emphasis on connectivity in landscapes and river networks illustrates the importance of conceptual models and methods of inquiry in governing the questions that scientists ask. If we view rivers as complex systems with multiple interactions between different components, we are more likely to focus on the factors that control those interactions and on ways to quantify and predict the interactions. If we view rivers as predominantly physical systems, we are more likely to neglect the interactions among hydraulics, sediment dynamics, and aquatic and riparian organisms. Even when not explicitly recognized, our conceptual models of rivers tend to constrain the questions that we consider interesting and important and the methods we use to examine these questions (Grant et al., 2013). Studies of sediment transport, for example, that employ a *Eulerian* framework focus on the flux of sediment within a spatially bounded area—a very useful approach for developing a sediment budget, whereas a *Lagrangian* framework in which specific objects are tracked through time can provide more insight into actual mechanisms of sediment movement (Doyle and Ensign, 2009).

A conceptual model results from assumptions about how a river functions. The conceptual model can be qualitative or quantitative. A quantitative model can be more precise than a qualitative model, but is not necessarily more accurate. Drawing on the second chapter of Leopold et al.'s (1964) fluvial geomorphology text for inspiration, the remainder of this section uses a landscape with which I am very familiar to explore the different conceptual models and approaches that investigators employ to understand river segments, river networks, and the greater landscape.

1.5.1 *The Colorado Front Range*

Atop the Precambrian-age crystalline rocks that form the continental divide in Colorado, you can stand shivering in the cold wind even at the height of summer. Here, 4000 m above sea level, bedrock topography crests in a series of ridges and peaks that divide water flowing west to the Pacific Ocean and water flowing east to the Atlantic (Figure 1.2). In some places the divide is a sharp-edged ridge of

Figure 1.2 Landscapes and river corridors in and adjacent to the Colorado Front Range. Upper left: view east from the summit surfaces at the continental divide. The coarse blocks in the foreground are periglacially weathered boulders and bedrock. The surface drops steeply into a glaciated valley that transitions downstream (out of sight) into a fluvial valley. Upper right: view northwest from a hogback, an asymmetrical hill of sandstone and limestone strata dipping steeply to the right in this view, with an intervening valley formed in shales. Lower right: the South Platte River near Fort Morgan, Colorado, in the low relief environment of the Great Plain. This sand-bed channel was historically much wider and had a braided planform, but flow regulation has resulted in encroachment of riparian vegetation and transformation to a single relatively narrow channel. This river heads high in the mountains. Lower left: view of smaller drainages that head on the Great Plains, here at Pawnee National Grassland. These channels have downcut within the past few decades, largely via piping erosion.

bedrock and periglacial boulders with talus chutes and waterfalls. In other places, small alpine streams meander across broad, gently undulating surfaces.

Sharp or broad, the heights drop precipitously down to glacial cirques and troughs. Rivers alternate between paternoster lakes and steep cascades as they flow through subalpine conifer forests. Beyond the terminal glacial moraine, each valley continues downward, alternating between steep, narrow gorges in which the river flows turbulent and aerated or relatively wide canyons with gentler gradients along which the river flows through pools and riffles.

Climate grows progressively warmer and drier at lower elevations, and subalpine forest gives way to more open montane forest with more frequent wildfires and associated debris flows. Warm, moist masses of air moving inland from the southeast during summer are forced upward as they near the Colorado Rockies, and the water vapor being transported with the air masses cools, condenses, and falls as rain. Most of this moisture is wrung from the clouds at the lower to middle elevations of the mountains, which can experience flash floods from convective storms, as well as the late spring snowmelt floods that flow down from the highest portions of the river network each year.

At the base of the mountains on the eastern side, the rivers gradually change from boulder- to cobble-bed channels as they flow through a series of steeply tilted sedimentary rocks forming asymmetrical hills. Beyond the hills lies the gently undulating topography and steppe vegetation of the semiarid Great Plains, where sand-bed channels shrink back to a trickle after the annual snowmelt peak flow.

The dramatic topography and strong elevational contrasts in climate and vegetation dominate initial impressions of the Colorado Front Range. This leads to questions about how river process and form change moving downstream, and what factors influence these changes. At a basic level, we can address these questions using empirical or theoretical approaches. *Empirical* approaches are largely inductive. In logic, to induce is to conclude or infer general principles from particular examples. In an empirical approach, data are collected and analyzed in order to establish relationships between variables. A fundamental challenge to empirical understanding of rivers lies in generalizing from empirical results defined by using a restricted database. If I measure bedload transport along a cobble-bed mountain river segment for a year and demonstrate that the majority of transport occurs when flow equals or exceeds half of the bankfull depth, can I extrapolate from this site to other rivers? What if I repeat the measurements on a sand-bed river of the plains and find that bedload transport begins at a much lower level of flow?

Theoretical approaches formulate and test specific statements based on established principles. To deduce is to reason from the general to the particular. Theoretical approaches are more deductive, but are typically hampered by a relative lack of established geomorphic theory. Consequently, theoretical approaches to river process and form commonly draw heavily on related fields such as hydraulic engineering in which the theory represents a system much more simple than most natural river channels.

Theoretical approaches to bedload transport developed by hydraulic engineers, for example, assume that bedload transport (i) begins once flow energy exceeds a critical level defined by the average grain size of the sediment and (ii) is proportional to the level of excess energy beyond the critical energy. The second assumption is illustrated by a generic equation for bedload transport rate q_b

$$q_b = k(\tau - \tau_c)^n \tag{1.1}$$

where k is an empirical constant, τ is boundary shear stress, τ_c is critical boundary shear stress for entrainment, and n is an empirically derived exponent. This equation implies that bedload transport is proportional to the amount of shear stress above the critical level for moving sediment. Equation 1.1 is an example of a *flux equation*. For rivers, flux equations usually refer to flow–sediment interactions and processes such as sediment flux within a channel

The relatively narrow grain-size range of sand-bed channels makes it easy to specify average grain size, and the relative ease of mobility of sand grains makes the second assumption above reasonable. In a cobble- or boulder-bed channel, however, the wider range of grain sizes means that larger grains can shield smaller grains from the force of the flow and limit the movement of the smaller grains. Consequently, average grain size may not be a particularly useful parameter for specifying the start of bedload transport. Larger grains at the streambed surface can prevent the movement of underlying smaller grains and create turbulence, so that bedload transport is not likely to have a linear relationship with flow energy.

The problem of characterizing bedload transport in mountains and plains rivers can also be described using the dichotomy of deterministic versus probabilistic. *Deterministic* approaches assume that physical laws control river process and form. Once these laws are known, river behavior can be predicted for a given set of conditions.

Deterministic modeling of river processes relies on five basic equations:

- continuity equations for water and sediment,
- the flow momentum equation,
- a flow resistance equation, and
- a sediment transport equation.

Conservation equations or *continuity equations* are based on the fact that mass, momentum, and energy cannot be created or destroyed in any process. The continuity equation for flow is simply

$$Q = w\, d\, v \tag{1.2}$$

where Q is discharge, w is flow width, d is flow depth, and v is mean velocity. An example of a sediment version is the Exner equation for sediment continuity,

$$\left(1 - \lambda_p\right) \frac{\partial \eta}{\partial t} = -\frac{\partial q}{\partial x} \tag{1.3}$$

where λ_p is bed porosity, η is bed elevation, t is time, q is volume transport rate of bed material load per unit width, and x is direction of flow (Parker et al., 2000). Another example of a continuity equation is a sediment budget that equates sediment storage to sediment input minus output.

The flow momentum equation is based on Newton's second law of motion and is well defined theoretically. Momentum is a vector defined by the product of mass and velocity. Momentum per unit time of water in a channel is $\rho Q v$, where ρ is water density, Q is discharge, and v is average velocity (Robert, 2003).

The flow resistance and sediment transport equations used in deterministic modeling of river processes will include empirically derived coefficients. Deterministic modeling can thus use both empirical and theoretical understanding of a system, but assumes that river process and form can be directly predicted based on knowledge of existing parameters.

As the particular component of a river system being modeled increases in complexity, the interactions are increasingly difficult to represent using a set of closed equations, and predictions become less reliable (Knighton, 1998). *Probabilistic* approaches reflect an assumption that natural systems are so complex that complete deterministic explanations are unrealistic because natural systems include inherent randomness. The ability to specify appropriate empirical flow resistance and sediment transport coefficients in boulder-bed mountain streams, for example, is limited by the extreme spatial variability in bed grain size, as well as irregularities in cross-sectional geometry caused by pieces of wood and lateral constrictions from bedrock outcrops or very large boulders. Under these circumstances, it is more effective to acknowledge a substantial level of uncertainty in predicting bedload transport: bedload movement may be described as occurring when discharge falls within upper and lower bounds, rather than as a direct relationship between discharge and bedload transport.

Another approach to predicting bedload transport would be to use a force equation. *Force equations*, typically the balance of forces involved in erosional and depositional processes, describe a critical level beyond which a process such as movement of sediment on the streambed or erosion of the stream bank begins. An example of a simple force equation for entrainment of a sediment particle in a river is

$$\tau = \gamma_f D S \tag{1.4}$$

where τ is shear stress acting on the sediment, γ_f is unit weight of the fluid, D is flow depth, and S is water-surface slope (Andrews, 1980). Again, the less spatial and temporal variation there is in a system—think sand bed (relatively uniform grain sizes) rather than boulder bed—the simpler it is to specify the forces at work and to accurately assign average values to parameters such as flow depth and water-surface slope in Equation 1.4.

Because natural rivers are commonly quite spatially and temporally variable, geomorphologists try to simplify process and form using *physical experiments* in which one or more variables are directly and systematically manipulated to observe the effect on the whole system. Such manipulations are typically conducted in a laboratory setting or, more rarely, in the field.

Bedload transport in the boulder-bed mountain rivers of the Colorado Front Range occurs 24 hours a day during the snowmelt peak. Much of the transport actually occurs in the early hours of the morning when the previous afternoon's snowmelt runoff comes down the river. Instead of attempting to directly measure bedload movement, and perhaps missing some of the sediment movement by not sampling the entire channel width or sampling continuously, useful insights into sediment dynamics can be gained by creating a scaled-down river in a flume and then measuring changes in bedload transport as discharge is varied. Physical experiments present challenges of scaling forces (can you effectively simulate the turbulence and associated hydraulic forces of a flow several meters deep?) and of including all relevant variables (can you effectively simulate fluctuations in upland or tributary sediment supply to the main channel?). Experiments can nonetheless provide useful insights into process and form in real channels.

Rivers can also be investigated by developing *numerical simulations* in which those variables and interactions considered to be relevant are quantified (Coulthard and Van de Wiel, 2013). Simulation outcomes are then compared to real rivers to evaluate the accuracy of parameterization and, once such accuracy is established, to test scenarios such as the effect of altering water or sediment yields to a river. Numerical simulations can be based on some combination of theoretical and empirically derived equations that can be deterministic or probabilistic. A numerical simulation of bedload transport, for example, might specify channel geometry, streambed grain-size distribution, discharge, and sediment input from upstream, and then use an equation such as Equation 1.3 to predict bedload flux. Among the challenges of numerical simulation are identifying the relevant variables and processes and parameterizing these variables and processes.

In addition to downstream differences in streambed substrate and bedload dynamics, some of the more obvious changes along river networks in the Colorado Front Range are the transitions from alpine meadows, to relatively dense subalpine forest of spruce and fir, more open montane pine forest, and finally semiarid steppe. Along the forested portions of the river networks, wood recruited from riparian forests can strongly influence channel process and form. These interactions illustrate another commonly used conceptual model of rivers as complex and nonlinear systems.

A *complex system* is composed of interconnected parts that as a whole exhibit one or more properties, including behavior, not obvious from the properties of the individual parts. A complex system displays self-organization over time and emergence over scale. Emergence is defined as patterns that arise from a multiplicity of relatively simple interactions. A tree topples into a river, for example, with a portion of the roots still attached to the bank. The downed tree extends into the river, trapping smaller pieces of wood in transport and forming a logjam. The logjam blocks flow, creating a backwater of lower velocity where sediment in transport settles out. As the elevation of the streambed increases, flow begins to spill over the channel banks and erodes a secondary channel that branches away before rejoining the main channel downstream. Bank erosion during formation of the secondary channel undermines more trees that fall into the river, forming additional logjams that facilitate further channel branching. Eventually, a network of branching and rejoining channels that enhance wood recruitment and storage is present. The tree fall and its consequent effects that result in a multi-thread channel segment are an example of a complex system (Figure S1.5).

In a *nonlinear system*, output is not directly proportional to input such that, mathematically, the variable to be solved for cannot be written as a linear combination of independent components because of interactions among the components. For example, pieces of wood floating downstream in a river are influenced not only by the hydraulic force of the flowing water, for example, but also by the movement of adjacent pieces of wood or the trapping effect of stationary instream wood. Because the movement of wood down the channel does not depend only on hydraulic force, this movement is an example of a nonlinear system.

Although phrases such as nonlinear and complex systems were not commonly used until the 1990s, the behavior described by these phrases was recognized decades earlier in descriptions of river process and form within the work of G.K. Gilbert (1877) and, in the mid-twentieth century, the work of Luna Leopold, Stanley Schumm, and others (Supplemental Section 1.5).

Rivers are also viewed as *open systems*, characterized by a continual exchange of matter and energy with the surrounding environment (Chorley, 1962). Such exchanges might be obvious at the scale of a channel segment with fluxes of water and sediment *from* upstream and upland sources and *to* downstream portions of the river. As emphasized in the opening discussion of connectivity, even the largest river networks also experience continual inputs of matter and energy from the atmosphere and the lithosphere, sometimes accomplished by the activities of organisms. Snow falling in the Colorado Front Range reflects the dynamics of cold, dry Arctic air masses moving southward and interacting with slightly warmer air carrying much more moisture and moving inland from the Pacific Ocean. The melting of the resulting snowpack, and consequently the timing and volume of snowmelt runoff in the rivers, is influenced not only by air temperature, but also by deposition of wind-blown dust that can come from nearby sources such as the deserts of the southwestern United States, and also from very distant sources in Asia (Painter et al., 2010).

Viewing a river as a complex, open system implies that, at whatever scale the river is considered, it contains multiple, interacting components. Interactions between components include feedbacks, thresholds, and lag times, and can create equifinality.

Feedback refers to interactions among variables. Self-enhancing feedbacks promote continuing change, as when sand grains saltating across bedrock are preferentially trapped on accumulations of sand that increase with time. The fallen log that initiates a logjam and eventually a network of secondary channels that promote additional wood recruitment and logjams is another example of a self-enhancing feedback. Self-arresting feedbacks limit change. For example, a lateral channel constriction causes an increase in flow velocity, which results in erosion of the constriction until the velocity drops below a magnitude capable of causing erosion of the channel boundaries.

Thresholds involve abrupt changes in process or form. External or extrinsic thresholds are crossed as a result of changes in external controls. Internal or intrinsic thresholds can be crossed in the absence of changes in external variables (Schumm, 1979). An example of an external threshold comes from hillslope hydrology, when the early stages of precipitation cause shallow infiltration and relatively slow downslope movement of water via subsurface diffusion. When sufficient water infiltrates to reach deeper layers with preferential flow paths in the form of soil pipes, downslope water delivery to channels abruptly becomes much more rapid. In this example, the abrupt change in downslope water delivery is externally forced by increasing volumes of precipitation.

The ephemeral tributaries that head on the dry eastern steppe of Colorado provide an example of internal thresholds. Over hundreds to thousands of years, these channels alternately incise to form steep-sided gullies or arroyos, and then aggrade to form relatively shallow swales (Figure 1.2). These alternating episodes of cut and fill can represent crossing of an external threshold in response to fluctuations in precipitation, vegetation, and runoff.

Alternating cut and fill can also occur in response to crossing an internal threshold of sediment transport within the channel. Stream flow in these channels is brief and infrequent, and sediment can be deposited midway down a channel as discharge declines because of evaporation and infiltration into the streambed. Repeated deposition of sediment partway along the stream's longitudinal profile can develop a steeper section of the bed at which a headcut eventually forms, triggering a wave of upstream-migrating incision. All of this can occur in the absence of any change in external variables such as precipitation, runoff, or sediment inputs.

Lag time typically refers to the delay between a change in an external variable, such as an increase in water yield, and the response of the river, such as bank erosion. The cobble-bed streams of the subalpine forest in the Front Range provide an example. Commercial ski resorts in this region divert river water to make snow for their ski runs. When this artificially created snow melts the next spring, the runoff commonly goes into a different channel than the source of the water. These receiving channels can have peak flows more than 200% larger than would result from natural runoff. Channels along which dense riparian vegetation and cohesive silt and clay increase bank resistance take longer to respond to increased peak flows than channels with less erosionally resistant banks (David et al., 2009) (Figure S1.6).

Where an external disturbance is very intense or widespread, lag times can be minimal. An intense wildfire in the montane zone of the Colorado Front Range during summer 2012 completely consumed hillslope vegetation over hundreds of hectares of pine forest underlain by weathered granite. The first rainstorms following the fire resulted in widespread erosion of hillslopes and headwater channels, and aggradation of larger channels (Figure S1.7).

Sediment accumulation in the larger channels is of particular concern because these rivers supply municipal drinking water to communities along the base of the Front Range. Water managers trying to maintain storage capacity at intake structures and limit turbidity associated with suspended sediment and organic matter could use the force equation for sediment entrainment, the continuity equations for flow and sediment, and the flux equation for bedload transport mentioned earlier to quantify sediment transport. They could also use *diffusion equations*, which describe the movement of matter, momentum, or energy in a medium in response to some gradient, such as the turbulent mixing of suspended sediment driven by gradients in flow energy (Robert, 2003). An example particularly relevant to the deposition of hillslope sediment mobilized after wildfire comes from unit sediment flux q in a river depositional system

$$q = v\frac{\partial h}{\partial x} \tag{1.5}$$

where v is diffusivity, h is elevation, and x is distance downstream (Voller and Paola, 2010).

Equifinality, also known as convergence, refers to the fact that different processes and causes can produce similar effects. This condition makes it difficult to infer processes from form (Chorley, 1962). Channel incision leading to terrace formation along the primary rivers of the Front Range might have resulted from (i) lowering of base level, or (ii) fluctuations in water and sediment supply to the river associated with the advance and retreat of Pleistocene valley glaciers, or with widespread deforestation and mining during the nineteenth century (Schumm, 1991). Data on the age, spatial extent, and stratigraphy of the terraces as well as independent information on the timing and nature of base level change, glaciations, and historical land use are necessary to explain terrace formation.

Underlying conceptualizations such as feedbacks and thresholds is one of the most widely used fluvial conceptual models: the idea that a river can exhibit various forms of stability, or equilibrium (Gilbert, 1877). *Equilibrium* typically refers to a condition with no net change, and is thus very dependent on the time span being considered. A river that undergoes substantial channel change during a short duration flood can nonetheless be in equilibrium when considered over a decade because subsequent smaller flows rework the erosional and depositional features created by the flood. Consequently, different forms of equilibrium can be distinguished with respect to time span (Figure 1.3).

Equilibrium implies that multiple interacting variables within a river can reach a state of stability. This is reflected in the widely used definition of a *graded river* as a channel in which streambed slope is adjusted to prevailing water and sediment discharges, such that the channel neither aggrades nor degrades and the slope remains constant over the time interval of interest (Mackin, 1948).

Equilibrium also implies that a river will change in response to changes in the supply of energy or material. Pleistocene valley glaciation in the upper portion of the Front Range changed water and sediment yields to downstream portions of the river network. Thinking of these river networks within a framework of equilibrium raises questions regarding how, and how rapidly or over what time span, the rivers responded to altered water and sediment supplies during glacial advance and retreat. One way to asses this is to evaluate downstream

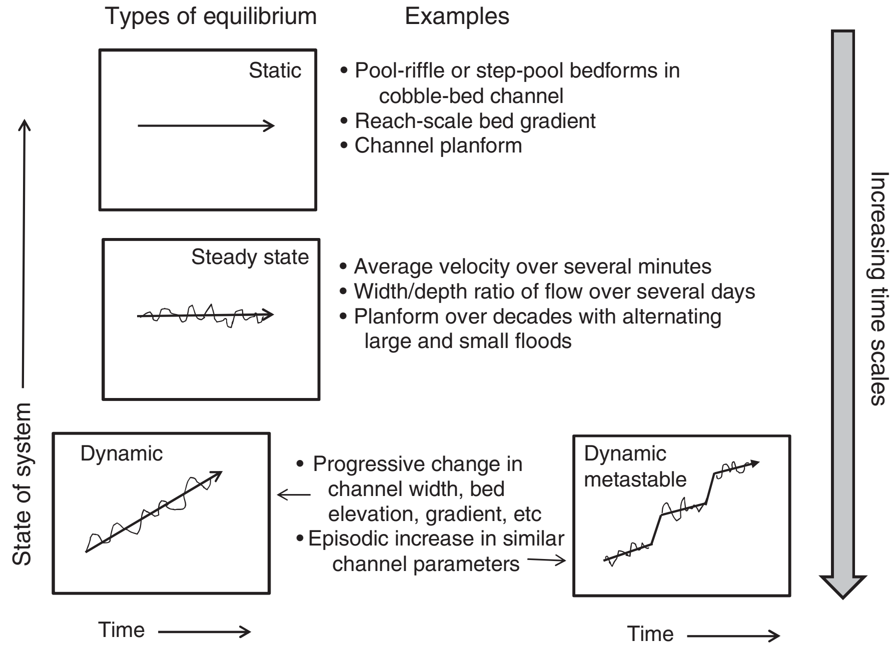

Figure 1.3 Schematic illustration of different types of equilibrium. Over the shortest time intervals, any particular variable representing state of the river system (e.g., channel planform or gradient) is *static* and unchanging. At progressively longer time intervals, the variable may be in *steady state*, with fluctuations about a consistent mean. At the longest time intervals, the mean value of the variable is likely to change, either progressively through time in *dynamic equilibrium* or in a stepped manner that reflects the crossing of thresholds, as in *dynamic metastable equilibrium*. The latter two cases are not, strictly speaking, equilibrium because the system exhibits net change over the time span being considered. The phrases are, however, widely used in this manner.

hydraulic geometry (Leopold and Maddock, 1953) relations for rivers in the Front Range. *Downstream hydraulic geometry* relations are empirical equations in the form of *power functions* derived from linear regressions of log-transformed data. These equations relate dependent variables of channel geometry to the independent variable of discharge. For example,

$$w = aQ^b \tag{1.6}$$

where w is channel width, Q is discharge, and the coefficient a and exponent b are determined from the linear regression.

Equation 1.6 implies that discharge is the primary influence on channel width. One implication is that values of channel width in the Front Range have fluctuated through time as the advance and retreat of valley glaciers altered discharge downstream from the glaciers.

A river in equilibrium is expected to have well-developed downstream hydraulic geometry such that variations in discharge explain most of the observed downstream pattern of variation in width (Wohl, 2004b). Headwater rivers within the glaciated portion of the Front Range exhibit less well-developed downstream hydraulic geometry relations, as indicated by lower values of the regression coefficient for w-Q regressions, than headwater rivers at lower elevations beyond the extent of Pleistocene glaciations. This suggests that rivers in the glaciated zone are still adjusting, more than 10,000 years after glacial retreat, to local variations in gradient, substrate resistance, sediment supply, and other factors that are affected by glaciation and that can influence channel width.

These rivers may be farther from equilibrium than otherwise analogous channels at lower elevations in the mountain range.

Equilibrium, or its absence, can also be described in terms of steady-state versus transient landscapes. A *steady-state landscape* can be defined with respect to denudation and topography as a landscape in which erosion and rock uplift are balanced such that a statistically invariant topography and constant denudation rate are maintained over a specified time interval (Whipple, 2001). A steady-state landscape thus exhibits equilibrium between uplift and erosion. *Transient landscapes* are those experiencing relatively brief (on a geological timescale) increases in erosion rate in response to, for example, active tectonic uplift (Attal et al., 2008). Ongoing change indicates that a transient landscape has not yet reached equilibrium following an external perturbation.

Exhumation of the Denver Basin at the eastern margin of the Colorado Front Range within the last few million years caused relative base level fall for the major rivers of the Front Range. Base level fall triggered a wave of incision that has been migrating upstream at an estimated rate of 0.15 mm/year (Anderson et al., 2006c). The location of contemporary active response to base level fall appears as a steepening—either a waterfall or steep section of rapids—in the longitudinal profile of each river. Portions of the river network upstream and downstream from this steeper zone are presently in steady state with respect to the base level fall, whereas the gradient of the steeper portion of the longitudinal profile is transient.

Contrasting river process and form between different portions of a region such as the Front Range underlies another approach to understanding rivers. Data for understanding rivers can be obtained from direct measurements in a field setting or from remote sensing imagery (Oguchi et al., 2013). Because of the long temporal scales over which river processes such as development of drainage networks or longitudinal profiles act, ergodic reasoning is also commonly used. *Ergodic reasoning* substitutes space for time by comparing features in different stages of development, under the assumption that variables other than time remain relatively constant. For example, drainage networks developed on otherwise comparable basalt flows of widely differing ages within a limited region can be compared to examine network development through time. The challenge of ergodic approaches is that variables other than time likely differ between sites being compared. Even if the basalt flows are identical in composition, for example, fluctuations in climate through time might cause the older networks to represent at least preliminary development under a climate different than that acting during development of networks on younger basalt flows.

Returning to the example of instream wood in forested streams of the Front Range, one way to investigate the importance of forest stand age on river–wood dynamics is to compare otherwise analogous stream segments flowing through forests of diverse age. Study design can be challenging: ideally, all other important parameters—drainage area, stream flow, sediment supply, valley geometry, and so forth—are similar between the stream segments, and only the age of the riparian forest varies. Comparisons using this ergodic approach suggest that old-growth forests with average tree age greater than 200 years have more instream wood, larger wood pieces, more closely spaced channel-spanning logjams, and consequently greater abundance of secondary channels and greater channel–floodplain connectivity (Beckman, 2013).

Stepping back to consider the river networks of the Front Range at a regional scale, many of the questions posed by Leopold et al. (1964) remain highly relevant today:

- What factors control hillslope process and form, and the initiation of channelized flow?
  - What is the rate of bedrock weathering and regolith production?
  - How and how rapidly does regolith move downslope into channel networks?
  - What variables influence the location of channel heads?
  - What processes result in the formation of channel heads?

- What factors govern the longitudinal profile of the rivers?
  - In particular, what is the relative importance of landscape-scale denudation in response to continuing adjustment to uplift, Pleistocene glaciations, and Quaternary relative base level fall?
  - What is the relative importance of longitudinal variations in bedrock erosional resistance, sediment supply, and flow regime?

Examining rivers in the context of the greater landscape, we can also add a series of new questions. Examples include:

- How do diverse types of connectivity vary throughout these river networks?
  - Are the alpine summit surfaces storing periglacial sediment, for example, or are they strongly coupled to adjacent glaciated valleys?
  - Channel–floodplain and channel–hyporheic-groundwater connectivity increase within lower gradient, wider valley segments, and then decrease in steep, narrow segments. What are the specific processes governing these downstream variations in connectivity?
- What are the magnitude and extent of human alteration of river networks?
  - When people of European descent settled the Front Range during the nineteenth century, they initiated lode and placer mining, extensive deforestation, and widespread flow regulation in the form of dams and diversions. Some of these activities ceased a century ago. Do river process and form still differ between networks in which these historical activities occurred and networks that were not altered in this way?
  - How does contemporary flow regulation alter physical and ecological functions of Front Range rivers?
  - Warming climate is resulting in changes in precipitation, soil moisture, wildfire regime, outbreaks of native insects that kill trees, and forest blowdowns. How do these pervasive alterations of forest dynamics and precipitation-runoff-stream flow influence channel process and form?
- What river-related geomorphic processes can be quantified in a manner applicable to diverse landscapes?
  - In an influential paper, Dietrich et al. (2003) highlighted the importance of developing *geomorphic transport laws* (GTLs) in the form of mathematical statements derived from a physical principle or mechanism, which express the mass flux or erosion caused by one or more processes. In order to be useful, it is important that such laws can be parameterized from field measurements, can be tested in physical models, and can be applied over relevant spatial and temporal scales. Existing GTLs include those for soil production from bedrock, linear slope-dependent transport of colluvium, and debris flow and river incision into bedrock.
  - What fundamental processes in the Colorado Front Range can usefully be expressed via GTLs (Figure 1.4)?
  - What processes are not yet adequately described by such laws?
  - How can we integrate GTLs with quantitative measures of connectivity?
- What components of river process and form are significantly influenced by biota?
  - Beaver were much more abundant in the Colorado Front Range prior to the nineteenth century. Have historical reductions in beaver populations and beaver dams influenced rivers regionally, or only local segments of rivers?
  - How do instream wood volume and associated geomorphic effects differ between subalpine and montane forests, or between steep, narrow valley segments and wide, lower gradient valley segments?
  - The extent and species diversity of riparian vegetation differ markedly between steep, narrow valley segments and wide, lower gradient segments. How do these differences influence valley-bottom sediment storage, hyporheic and groundwater exchange, and water chemistry along Front Range rivers?

To quote Leopold et al. (1964, p. 18): "Partial explanations of these problems can be offered, but

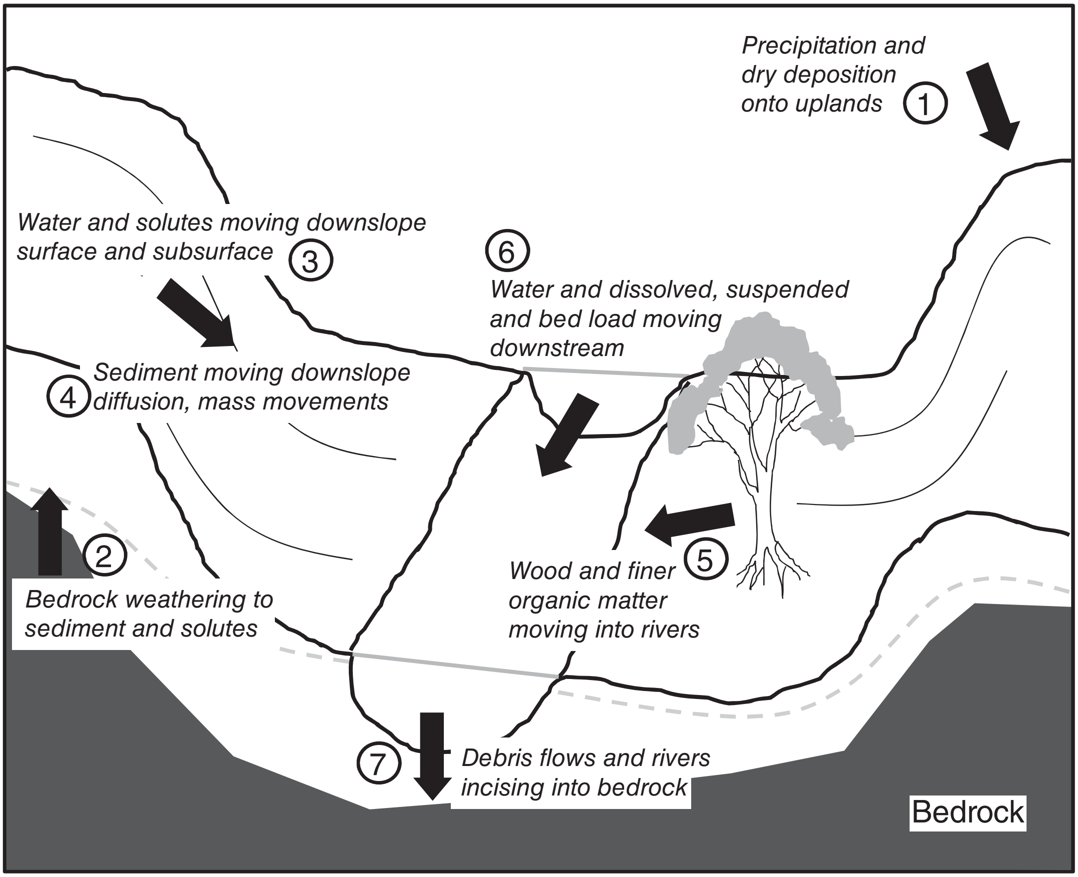

Figure 1.4 Schematic illustration of fundamental fluxes in any landscape. Among those for which some form of geomorphic transport law (GTL) has been proposed are 2 (bedrock weathering), 4 (downslope movement of sediment via diffusion), and 7 (incision into bedrock), although these GTLs require additional testing and parameterization for specific field settings. Empirical or theoretical equations have also been proposed for 1 (precipitation and dry deposition), 3 (downslope movement of water and solutes), and 6 (water and sediment movement downstream). Again, these equations require testing and parameterization. All of the equations proposed for these fluxes assume that average values can be quantified based on prevailing conditions. Using bedrock weathering as an example, chemical reactions are a function of factors such as temperature and the amplitude of temperature oscillations (Supplemental Section 2.1.2). Explicitly incorporating connectivity requires quantifying variations in prevailing conditions through time that limit or enhance fluxes, such as short-term variations in weather and longer-term variations in climate that influence temperature and thus chemical weathering rate.

more complete explanations require much more knowledge of processes than is presently available."

1.6 Only connect

E.M. Forster took "only connect" as the epigraph for his novel "Howard's End." Forster referred to connections between individual people and different classes within a society, but this phrase is also particularly apt for understanding rivers. If we can extend our understanding sufficiently and

- connect rivers to landscapes
- connect contemporary river configuration to human and geological history
- connect site-specific river characteristics to universal river process and form

- connect field observations to physical experiments and numerical simulations, and
- connect geomorphic knowledge of river process and form to
  - ecological knowledge of aquatic and riparian organisms,
  - geological knowledge of rock substrates and tectonics,
  - social science knowledge of human history and decisions regarding resource use, and
  - biogeochemical knowledge of aqueous chemistry

… then we will be making progress in understanding the complex and fascinating world of rivers (Figure S1.8).

Chapter 2
Creating channels and channel networks

2.1 Generating water, solutes, and sediment

2.1.1 Generating water

Rain is the most common form of precipitation (Barry and Chorley, 1987). Precipitation includes all liquid and frozen forms of water—rain, snow, hail, dew, hoar frost, fog drip, and rime—moving from the atmosphere toward the ground surface, but rain and snow constitute by far the greatest contributors to precipitation worldwide. Rainfall at a measurement site is typically characterized in terms of maximum or average intensity (amount per unit time), duration, total amount, spatial extent of a rainstorm, and frequency of rainstorms.

The major types of precipitation can be differentiated into convective, cyclonic, and orographic, based on the primary mode of uplift of the air that causes water vapor to condense into water droplets or ice crystals. Supplemental Section 2.1 describes the characteristics of each type of precipitation, as well as global patterns of precipitation.

The characteristics of precipitation strongly influence the amount of water reaching channel networks and the pathways that water follows downslope into channels. Most river networks are influenced by persistent types of precipitation, such as predominantly convective storms, although moderate- to large-sized networks commonly experience multiple forms of precipitation, such as winter snowfall and summer convective storms. Supplemental Section 2.2.1 describes human influences on precipitation in the context of warming climates.

Precipitation reaching Earth's surface can be stored for periods of hours to months in a solid form as snow, or for decades to thousands of years as glacial ice. Melting of this solid, stored water strongly influences rivers at high altitudes and latitudes. The relative importance of glacier melt, snow melt, and rainfall typically vary by latitude and location with respect to atmospheric circulation patterns and, within a region, by elevation. Precipitation and runoff patterns within the Himalayan massif exemplify the effect of elevation. Up to 60%–80% of river discharge originates as monsoon rainfall at

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lower elevations on the southern side of the massif, whereas glacier and snow melt contribute 50%–70% of river discharge at higher elevations and on the northern side of the massif (Gerrard, 1990; Wohl and Cenderelli, 1998).

Worldwide, glacier and snow melt are progressively more important at higher latitudes and higher elevations, and the seasonal melt contribution is delayed later into the summer with increasing latitude and elevation. As the proportion of the basin area covered by ice and snow increases, progressively more of the total runoff occurs during summer (Chen and Ohmura, 1990; Collins and Taylor, 1990). Interannual variation in runoff also declines with greater snow and ice coverage, although the relation is not linear (Collins, 2006b). Human influences on glacier and snow melt in the context of warming climates are discussed in Supplemental Section 2.2.1.

2.1.2 Generating sediment and solutes

The bedrock underlying a drainage basin is the starting point for much of the sediment and solutes that eventually move downslope and into rivers, although both particulate and dissolved materials can also enter rivers via eolian transport and wet and dry atmospheric deposition (e.g., Schuster et al., 2002; Galloway et al., 2004). An idealized vertical profile through the weathering zone (Figure S2.1) has unweathered bedrock at the base, overlain by *regolith*, defined as rock that is weathered to any degree. Regolith is subdivided into *weathered rock* that is fractured and/or chemically weathered but has not been mobilized by hillslope processes or bioturbation, overlain by *saprolite*, which retains the original rock structure yet has been sufficiently altered that it can be augured through or dug with a shovel, and *mobile regolith* (Anderson and Anderson, 2010). Mobile regolith includes *soil*, which is organized into horizons by soil-forming processes. In practice, mobile regolith and soil are commonly used interchangeably, as in this chapter.

At the regional scale, lithology, tectonics, and climate influence processes and rates of bedrock weathering. Minerals that crystallize from molten material at high temperatures are typically less resistant to weathering than minerals such as quartz that crystallize at lower temperatures, so the mineralogical composition of bedrock influences weathering (Anderson and Anderson, 2010; Ritter et al., 2011). Tectonic stresses and regional-scale deformation can fracture bedrock at various depths in the crust, increasing surface area and making the rock more susceptible to chemical alteration and to other physical weathering processes such as freeze–thaw cycles (Anderson and Anderson, 2010). Chemical weathering results in chemical alteration of the regolith, and in dissolution of more reactive minerals, which release ions that are transported into ground and surface waters as solutes. Chemical weathering is facilitated by warm, wet conditions, whereas physical weathering is greatest in moderately wet climates with low temperatures that promote frost action (Ritter et al., 2011).

At smaller spatial scales, factors such as soil erosional flux, hillslope morphology, and biota influence rates of weathering. Bedrock must weather at a rate equal to or greater than the rate of erosion if soil is to persist, and many soil profiles appear to reach a steady-state thickness such that soil production is balanced by removal (Lebedeva et al., 2010). Chemical weathering rates, soil production rates, and hillslope curvature decrease with increasing soil depth (Heimsath et al., 1997; Burke et al., 2007). Consequently, any process that alters soil depth by moving soil downslope—mass movements, creep, bioturbation—also alters rates of bedrock weathering and soil production. Humans, in particular, move tremendous amounts of sediment globally (Hooke, 2000), resulting in long-term and spatially extensive changes in bedrock weathering and soil distribution and development (Montgomery, 2007). Hillslope morphology is closely connected to erosional fluxes. Upper, convex portions of a hillslope are likely to have steady removal of weathered products, for example, whereas lower, concave portions of the hillslope can accumulate weathered materials. At larger scales, hillslope curvature (and soil production) varies inversely with soil depth (Heimsath et al., 1997). Plants and animals influence rates of weathering by excreting organic acids, and physically disrupting weathered materials, as when plant roots expand into bedrock joints or burrowing animals churn the regolith. Supplemental Section 2.1.2 provides more details on soil production.

As noted in the first chapter, sediment production is rarely uniform across even a small catchment as a result of differences in lithology, climate, tectonics, biotic communities, hillslope morphology, or land use. Investigating the tectonically active, semiarid Western Transverse Ranges of California, USA, Warrick and Mertes (2009) found that about half of the suspended sediment discharge from the region came from a very small proportion of the landscape underlain by weakly consolidated bedrock and associated with the highest rates of tectonic uplift. Similarly, disproportionate sediment generation has been demonstrated for the Amazon and Mississippi River basins (Meade et al., 1990; Meade, 2007). Milliman and Syvitski (1992) estimate that mountain and high mountain (>1000 m elevation) drainages collectively cover 70% of the global land area, but contribute 96% of total river sediment yields. Analogous patterns also appear to govern solute production. Lyons et al. (2002) estimate that high-standing ocean islands in the southwest Pacific, for example, make up only about 3% of Earth's landmass, but contribute 17%–35% of particulate organic carbon entering the world's oceans annually.

2.2 Getting water, solutes, and sediment downslope to channels

2.2.1 Downslope pathways of water

The great majority of water flowing in a river passes over or through an adjacent upland and its regolith before reaching a channel (Kirkby, 1988). The pathways followed by the water entering a river exert a strong influence on the volume and timing of flow in the channel. Precipitation falling toward the ground does not necessarily reach a river, however.

Precipitation can be intercepted by plants and either directly *evaporate* from the plants or be taken up by the plant tissues and then released back to the atmosphere through *transpiration*. Evaporation and transpiration are strongly influenced by energy availability at the surface and water availability in the subsurface. Transpiration also reflects plant physiology (Kramer and Boyer, 1995) and the ability of plant roots to take up, redistribute, and even selectively extract water from different subsurface depths (Lai and Katul, 2000). Recent estimates suggest that transpiration represents the largest water flux from continents (i.e., larger than rivers), composing 80%–90% of combined evaporation and transpiration—commonly known as evapotranspiration (Jasechko et al., 2013).

Interception losses from evaporation and transpiration can reach 10%–20% beneath grasses and crops and up to 50% beneath forests (Selby, 1982). Interception can vary substantially during the course of the year in regions with seasons during which plants go dormant or during which soil moisture declines and plants store more water (Link et al., 2005). Plants can also directly intercept cloud water. In cloud forests, this interception can reach 35% of mean annual precipitation (Bruijnzeel, 2005). Plant stems can concentrate the movement of precipitation toward the ground via *stemflow*, which has been described for environments as diverse as tropical montane forest, semiarid loess terrain, and a seasonal cloud forest in which stemflow accounted for 30% of total precipitation (Hildebrandt et al., 2007).

Precipitation that does reach the ground surface can be evaporated from bodies of standing or flowing surface water, or from the ground surface. The pathways taken by precipitation reaching the ground surface are strongly dependent on the soil cover. Soil cover changes continuously in response to water movement and processes of weathering, bioturbation, sediment transport, and land use or other changes in land cover (Brooks, 2003). Development of crusts, compaction, and sealing (for example, with clay particles) decrease surface permeability (Slattery and Bryan, 1994; Kampf and Mirus, 2013), as does development of soil water repellency, or *hydrophobicity*, from burnt plant litter (Martin and Moody, 2001). Precipitation is equally dynamic, and the intensity and volume of precipitation reaching the land surface vary substantially through time and space. Consequently, downslope movement of water on the surface and in the subsurface is rarely in a steady-state condition (Wainwright and Parsons, 2002).

Water can flow down slope at the surface as *overland flow*, which is used to describe surface flow outside the confines of a channel. Overland flow includes *Hortonian overland flow*, also known as *infiltration excess overland flow* (Horton, 1945), if the infiltration capacity is low relative to precipitation intensity (Figure 2.1). Hortonian overland flow is most common where vegetation is sparse, slope gradients are steep, regolith is thin or of low permeability (e.g., clay), and precipitation intensities are high: characteristics of arid and semiarid regions. The great majority of most natural watersheds do not produce Hortonian overland flow, but human land uses that compact the soil or cause erosion of permeable, near-surface soil layers promote Hortonian overland flow. The most extreme example of this effect is paved, impermeable urban surfaces.

Overland flow can also move downslope as *saturation overland flow*, also known as *saturation excess overland flow* (Dunne and Black, 1970a, 1970b). This is a combination of direct precipitation onto saturated areas and return flow from the subsurface as saturation occurs. The moisture content of the regolith before, during, and after precipitation exerts a particularly important control on saturation

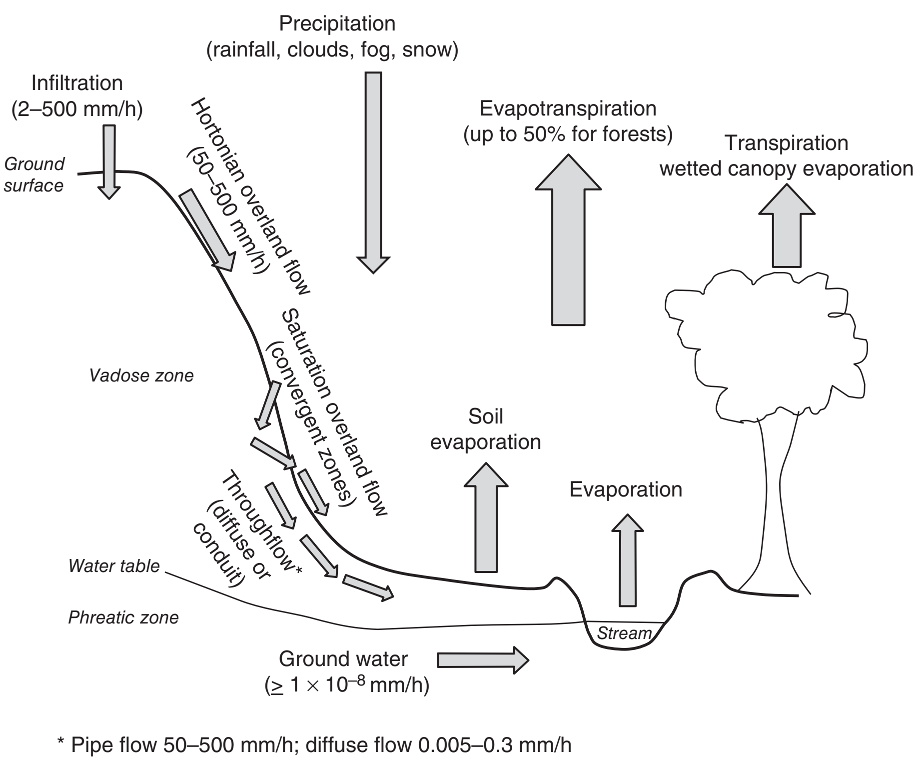

Figure 2.1 Schematic illustration of different types and rates of downslope water movement. Downward arrows reflect movement of water toward (precipitation), into (infiltration), across (overland flow) or beneath (throughflow, groundwater) Earth's surface. Upward arrows reflect water movement back into the atmosphere.

overland flow, which is influenced more by antecedent soil moisture and subsurface transmissivity than by slope steepness (Montgomery and Dietrich, 2002). As prolonged precipitation allows deeper and less permeable regolith layers to become saturated, subsurface flow expands to include areas progressively closer to the surface (Knighton, 1984). Conditions that favor saturation overland flow include high permeability near the surface, a humid climate with high cumulative water input, and gentler slopes with shallow soils that cannot drain as easily as steep slopes in which hydraulic gradients approximately parallel the steep surface topography (Kampf and Mirus, 2013) (Figure 2.2).

Saturation overland flow is rare outside of convergent flow zones such as hillslope concavities (Dietrich et al., 1992). Several scenarios can cause saturation flow at other points along a hillslope, however, including: topographic breaks, permeability contrasts (e.g., roads or pavement), low subsurface storage capacity, geologic structures that promote zones that saturate readily (e.g., layered basalt flows) (Mirus et al., 2007), exclusion from frozen soil, or snow melt over saturated soil (Kampf and Mirus, 2013). Saturation overland flow can also occur in tropical rainforests where a sharp drop in permeability with depth and high rain inputs lead to perched water tables (Bonnell and Gilmour, 1978; Elsenbeer and Vertessy, 2000).

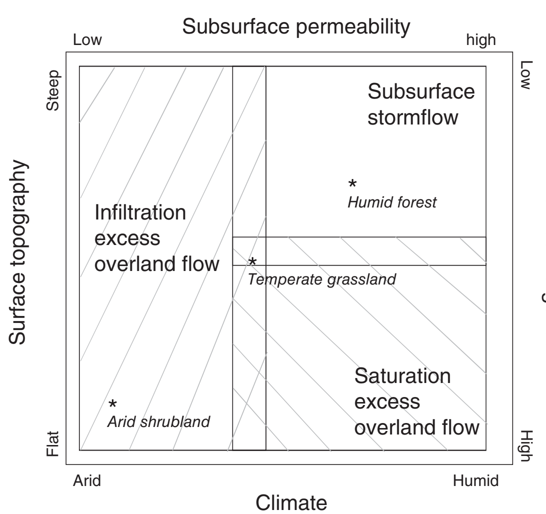

Figure 2.2 Environmental controls on dominant runoff generation mechanisms. Runoff generation mechanisms represented by patterned fields: overlapping fields indicate that no single runoff mechanism dominates hydrologic response. Asterisks represent approximate conceptualizations, for example environments (from Kampf and Mirus, 2013, Figure 10).

Overland flow commences when water ponds on the land surface to sufficient depth to begin flowing downslope. Surface roughness from microtopography and vegetation generate flow resistance that influences overland flow pathways (Bergkamp, 1998), particularly in humid environments with dense vegetation and organic litter, but the flow also modifies surface roughness. Vegetation also influences overland flow by altering infiltration. Shrub mounds in semiarid regions facilitate infiltration and generate less overland flow than intervening spaces, creating downslope pathways strongly coupled to vegetation patterns (Dunne et al., 1991).

Reported infiltration capacities from diverse locations range from 0 to 2500 mm/h (Selby, 1982). Infiltration can be extremely variable through space and time within a single small catchment because of the numerous factors that influence infiltration (Figure 2.3). The *variable source area* concept (Hewlett and Hibbert, 1967) reflects the fact that the area of a catchment actually contributing water to a channel extends during precipitation and contracts after the precipitation ends, with contributing area varying between 5% and 80% of the catchment (Dunne and Black, 1970b; Selby, 1982). Much of this variability likely reflects thresholds of activation for lateral subsurface flow (McDonnell, 2003).

Saturation overland flow typically occurs first in downslope portions of a catchment and then expands upslope (Dunne, 1978). In cold regions with low topographic relief, upslope areas can be the first sources of runoff as a result of varied local topography and water table position in relation to frozen ground (Spence and Woo, 2003). Spence and Woo (2006) use the *element threshold concept* to describe a catchment in which runoff is governed by the function of spatially and hydrologically

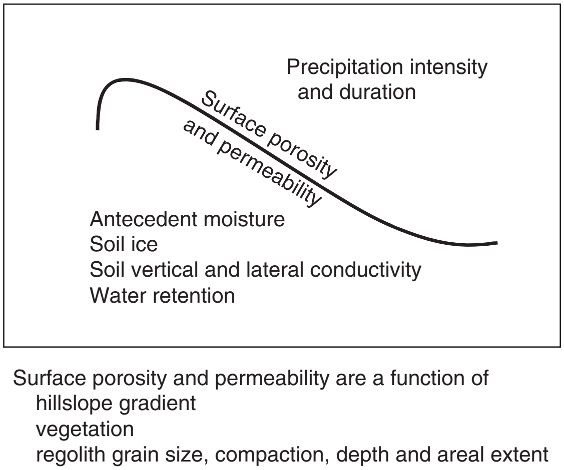

Figure 2.3 Schematic illustration of the variables influencing infiltration capacity. Line represents a side view of a sloping ground surface. Precipitation characteristics listed at upper right, surface characteristics listed along the line, and subsurface characteristics listed at lower left all help to create changes in infiltration capacity over small spatial and temporal scales.

distinct areas (elements), such as bedrock uplands, soil-filled valleys, and lakes, and the hydrologic linkages between elements. Each element can store, contribute, or transmit water, and the occurrence of these functions reflects the water balance of the element as well as connections to adjacent elements. The primary difference between the variable source area and element threshold concepts involves the assumed greater hydrologic connectivity within catchments operating under the variable source area concept.

Infiltrating water that remains in the subsurface can flow downslope in the vadose (unsaturated) zone above the water table as *throughflow*, or in the phreatic (saturated) zone below the water table as *groundwater*. In either case, subsurface water flows through interconnected voids between solid materials. Where the interconnected pores are small, flow through the porous medium is laminar and of low velocity (Kampf and Mirus, 2013) and is commonly described using Darcy's law (Darcy, 1856). When the void space in a porous medium is filled with water under saturated conditions, the *hydraulic conductivity*, or ease of water flow through the medium, is at its maximum value (Kampf and Mirus, 2013). When the void space is not completely filled with water, the connectivity of pore space and the hydraulic conductivity decrease.

Hillslopes are typically unsaturated at the ground surface and water movement is predominantly vertical during infiltration. Moisture is redistributed in the subsurface through both vertical and lateral movements that reflect the interaction of infiltration and percolation (wetting) with evaporation and transpiration (drying) (Kampf and Mirus, 2013). Water moving into the soil can propagate in the form of diffuse wetting fronts, fingered flow paths that result from wetting front instabilities, or preferential flow along conduits (Wang et al., 2003). Although lateral unsaturated flow can occur, lateral flow is most common under saturated conditions (Kampf and Mirus, 2013).

Diffuse throughflow depends on the general porosity and permeability of the unsaturated zone, whereas concentrated throughflow moves in preferential flow paths such as pipes or macropores (Dunne, 1980; Jones, 1981). *Macropores* are openings sufficiently large that capillary forces have an insignificant effect on the water running through the pores (Germann, 1990). Preferential flow within the macropore does not have time to equilibrate with slower flow through the surrounding matrix (Šimůnek et al., 2003) and is likely to be turbulent and thus not adequately described by Darcy's law. Lateral flow through macropores requires pore connectivity and a rate of water supply that exceeds the loss rate to the surrounding soil, which suggests that macropore flow is triggered at a threshold wetness level (Beven and Germann, 1982) (Figure S2.2). Once this threshold is exceeded, flow is much faster than diffuse, matrix flow. Rapid subsurface stormflow is particularly widespread on densely vegetated hillslopes in steep terrain where dense biological activity creates high concentrations of macropores, although the phenomenon has also been documented in semiarid climates (Kampf and Mirus, 2013).

Substantial preferential flow through macropores can facilitate the formation of soil pipes. *Soil pipes* are larger than macropores and are typically formed by subsurface erosion that enlarges animal burrows, root channels, or cracks from desiccation or unloading (Bryan and Jones, 1997). Pipes can be only a few centimeters in length and diameter, or 2 m in diameter and hundreds of meters long (Selby, 1982). Pipes tend to form just above a zone of lower porosity and permeability or along a cavity created by a burrowing animal or the decay of plant roots. Piping can occur in any region, but is particularly associated with drylands. In some catchments, piping can contribute up to nearly 50% of stormwater flow (Jones, 2010).

Pipe networks can exist at multiple levels in the regolith, with each level being activated by precipitation of different magnitude (Gilman and Newson, 1980; Kim et al., 2004). Rapid lateral flow via soil pipes or via discontinuities at the soil–bedrock interface appears to depend on thresholds, such that hillslopes "turn on" when sufficient water infiltrates (Uchida et al., 2001; McDonnell, 2003). This may help to explain the *old water paradox* in which pre-event water largely dominates storm runoff, suggesting that catchments store water for considerable periods of time but then promptly release the water during storms (Kirchner, 2003). The existence of thresholds also helps to explain abrupt changes in hydrologic response and water delivery to channels with different hillslope wetness states (Kampf and Mirus, 2013) (Figure S2.3).

Downslope water movement below the water table can also be quite complex as a result of spatial variations in depth and rate of movement of groundwater. The water table responds separately in riparian and hillslope zones, for example (Seibert et al., 2003), so that upslope area groundwater can be falling during the early portion of runoff while the riparian water table is rising. The hillslope groundwater can be slowly falling as part of the recession from rainfall several days earlier. Groundwater close to the stream is more likely to be in phase with runoff.

Other sources of complexity in groundwater dynamics include regional groundwater flow between watersheds (Genereux and Jordan, 2006). Small groundwater reservoirs such as those in mountainous regions can respond to seasonal processes such as snow melt runoff (McDonnell et al., 1998; Clow et al., 2003). Deep infiltration into bedrock can be important where bedrock is close to the surface or where highly porous and permeable or highly fractured bedrock can take up substantial volumes of water (Flint et al., 2008) (Figure S2.4). This type of deep subsurface flow can be reflected in slow or doubly peaked runoff response (Onda et al., 2001), in which the two peaks reflect unsaturated and saturated zone flow.

Subsurface flow typically dominates slopes with full vegetative cover and thick regolith (Dunne and Black, 1970a). Thicker soils increase the mean residence time of water on slopes and damp the temporal fluctuations of water movement in response to precipitation inputs (Sayama and McDonnell, 2009). Water moving downslope via Hortonian overland flow typically has the most rapid rate of movement (50–500 m/h), with progressively slower rates of movement during saturation overland flow, throughflow, and groundwater flow, which can move as slowly as 1×10^{-8} m/h (Selby, 1982). The exception to this generally slower movement from the surface to progressively deeper paths is concentrated throughflow in pipes or macropores, which can move quite rapidly (Figure 2.4).

The distribution of water among different downslope pathways can alter in relation to precipitation magnitude, intensity or duration during a storm, or on a regular annual basis in strongly seasonal climatic regimes. The dominant runoff processes also change with spatial scale (McDonnell et al., 2005). The manner in which moisture is released from pores within soil might condition runoff within a soil column, whereas partitioning between preferential and non-preferential flow governed by soil structure and rain intensity become more important at the plot scale. Spatial variation in soil depth strongly influences lateral water movement in transient subsurface saturated areas at the hillslope scale, and the proportion of water derived from different catchment geomorphic units such as hillslopes, riparian zones, and bedrock outcrops influence the river hydrograph at the catchment scale (McDonnell et al., 2005).

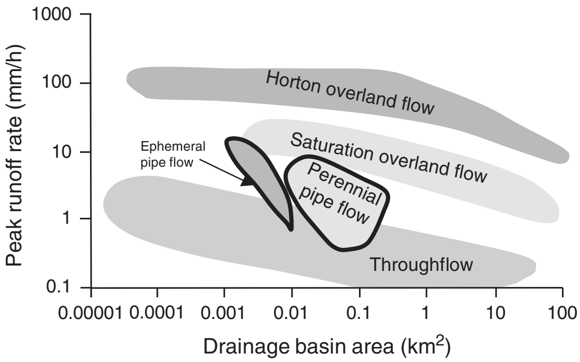

Figure 2.4 Generalized graph of hillslope hydrologic processes in relation to size of contributing area (from Jones, 2010, Figure 7b). The drainage basin area in this figure reflects total contributing area of the channel upstream from the point at which runoff enters a channel, but does not necessarily reflect the continuous extent of the area generating runoff. A 100 km^2 drainage basin area, for example, does not mean that Horton overland flow has to persist for tens of km before entering a channel.

Hillslope flow paths can be conceptualized as occurring along a spectrum from predominantly vertical to predominantly lateral (Elsenbeer, 2001). Lateral flow, in particular, is highly nonlinear and exhibits threshold behavior that influences downslope connectivity as small depressions in the underlying bedrock topography fill and spill or laterally discontinuous soil pipes and macropores self-organize into larger preferential flow systems as sites become wetter (Sidle et al., 2001; Hopp and McDonnell, 2009).

Karst terrains, cold regions underlain by permanently frozen ground, and the humid tropics form three distinctive subsets in terms of regional patterns of downslope pathways of water. *Karst* terrains have distinctive landforms and drainage as a result of greater rock solubility in the presence of natural water; typically, these terrains are associated with carbonate rocks and evaporites. Water flowing at the surface in karst terrains over rocks of low solubility can move very abruptly downward to the groundwater via *swallow holes*, which are open cavities on the channel floor where the channel flows onto carbonate rocks. Swallow holes can divert a portion or all of a surface river's discharge. Precipitation falling on karst terrains can percolate down to the groundwater via diffuse infiltration in the zone of aeration, a process known as *vadose seepage*. Precipitation can also move downward via a highly permeable zone—typically formed by the intersection of vertical joints in the bedrock or cylindrical solution openings—at the base of a closed depression, a process known as *vadose flow* or *internal runoff* (Ritter et al., 2011).

The *epikarstic zone* within the vadose zone is a heterogeneous interface between unconsolidated material and solutionally altered carbonate rock (Jones et al., 2003). The epikarstic zone is partially saturated and can delay or store and locally reroute vertically infiltrating water into the deeper, regional phreatic zone of the underlying *karstic aquifer*. Aquifers formed in karst terrains differ from other types of aquifers in that karstic aquifers typically have larger variations in porosity and permeability. Groundwater in these aquifers can move through intergranular pores within unfractured bedrock (*matrix permeability*), through joints and bedding planes created in the rock after deposition and lithification (*fracture permeability*), or through conduits with widths exceeding 1 cm that have been enlarged by solutional weathering (*conduit permeability*) (White, 1999).

The extent of frozen soil strongly influences downslope pathways of water in cold regions. In very cold regions, an *active layer* that thaws seasonally and ranges from 15 cm to 5 m thick overlies permanently frozen ground, or *permafrost*. Frozen soil impedes infiltration and limits percolation, with the result that a large portion of snow melt moves downslope as overland flow and is quickly delivered to channels (Vandenberghe and Woo, 2002). As the active layer thaws, the depth and importance of infiltration can change, but the presence of permafrost fundamentally limits deep infiltration and groundwater flow. Retreat of permafrost in response to global warming is creating major changes in downslope pathways of flow in cold regions.

Many catchments in the humid tropics have high runoff coefficients and quick hydrologic responses to precipitation. Infiltration rates in the humid tropics range from 0 to over 200 mm/h (Harden and Scruggs, 2003), however, and runoff generation reflects numerous influences, as in temperate environments (Scatena and Gupta, 2013). A variety of shallow subsurface flow paths contribute to the typically flashy response and high discharge per unit drainage area common in humid tropical catchments. Among these shallow subsurface flow paths are tension cracks that develop during the dry season and abundant macropores (Niedzialek and Ogden, 2005).

The downslope pathways of meltwater through snow and glacial ice are as complex as those through slopes of sediment and bedrock. Outflow from a glacier includes a base flow component supplied by groundwater discharge, runoff from storage zones within the ice, runoff from the firn water aquifer at the glacier's surface, and regular drainage from lakes (Gerrard, 1990). The melting of seasonal snow cover on the glacier and on non-glacial surfaces can produce an initial runoff peak, with subsequent melting of the glacial ice producing another runoff peak (Aizen et al., 1995). Runoff from the glacier can also vary during the melt season as channels develop on top, within, and beneath the glacier (Fenn, 1987; Nienow et al., 1998) and as air temperatures fluctuate (Hodgkins et al., 2009).

Runoff from snowpack depends on percolation times from the melt at the surface of the snowpack to the ground, as well as distance downslope from the melt source. Travel times through the snowpack dominate runoff in watersheds < 30 km^2 in the Sierra Nevada of California, USA, for example, whereas snowpack heterogeneity results in more consistent timing of peak runoff in watersheds > 200 km^2 (Lundquist et al., 2005).

Ice- and snow-covered portions of a catchment can sometimes retain meltwater and rainfall to a greater extent than snow-free portions of the catchment. The ability of the snowpack to retain or transmit rainfall partly depends on its structure because the presence of ice layers can substantially increase water retention compared to a homogeneous snowpack (Singh et al., 1998). The proportion of a basin covered by ice and snow typically varies through the seasons as the transient snow line shifts, resulting in seasonal variations in the magnitude of rainfall-induced runoff (Collins, 1998).

Supplemental Section 2.2.1 discusses measuring and modeling the downslope pathways of water, and describes human influences on downslope movement of water.

Although knowledge of downslope pathways of water has increased substantially during the past few decades, the fundamental questions remain: where does water go when it rains? What flow path does it take to streams? And, how long does water reside in the catchment? Knowledge of downslope pathways of water is critical to fluvial geomorphology because the resulting stream flow is the primary driving force in river networks.

2.2.2 Downslope movement of sediment

Sediment moves downslope through mass movements as slides, flows, or heave, and through gradual diffusive processes of rainsplash and overland flow. Diffusive processes typically involve individual grains rather than aggregates of grains. Whatever the transport mechanism, the portion of the soil profile actively transported downslope in steeper, forested terrains can reflect the rooting depth and consequent root wad thickness of fallen trees, which in turn reflects depth to the soil/saprolite boundary (Jungers et al., 2009).

Mass movements involve downslope transport of aggregates rather than individual particles and are typically strongly seasonal as a function of moisture availability and freeze–thaw processes (Hales and Roering, 2009). Mass movements are categorized based on the characteristics of the moving mass into slides, flows, and heave, each of which results from a decrease in the shear strength of the material or an increase in the shear stress acting on the material (Carson and Kirkby, 1972) (Figure S2.5).

A *slide* occurs when a mass of unconsolidated material moves without internal deformation along a discrete failure plane that can be curved and produce the rotational movement of a slump, or can be relatively straight. Slides typically result from a decrease in the shear strength of the soil as a result of weathering, increased water content, seismic vibrations, freezing and thawing, or human alterations such as deforestation and road construction. An increase in shear stress caused by additions of mass or removal of lateral or underlying support can also trigger a slide. Slides typically transition downslope into *flows*, which occur when the moving mass is sufficiently liquefied or vibrated to create substantial internal deformation during downslope movement.

Mass movements can also begin as flows initiated by runoff-dominated processes that cause progressive sediment entrainment or by infiltration-dominated processes that trigger discrete failures (Cannon et al., 2001). Mass movements resulting from runoff occur when sheetwash and rills entrain progressively more sediment downslope until the runoff concentrates in gullies and channels, eroding additional sediment from these conduits. This succession of events can occur during intense rainfall in arid environments, or after a wildfire that removes surface vegetation and plant litter, as described by Wohl and Pearthree (1991) for burned sites in Arizona, USA, (Figure S2.6) and Meyer and Wells (1997) for burned sites in Wyoming, USA. Infiltration can initiate local slope failures in the form of a slide that entrains more material as it moves downslope. This can occur during intense precipitation or rapid melting of a snowpack, or following a wildfire. Mass movements can also result from what Johnson and Rodine (1984) termed the "firehose effect," in which concentrated peak discharge at the outlet of a fan or the base of a steep, bedrock chute can mobilize accumulated sediment.

Abrupt mass movements recur frequently—almost annually—in many high-relief terrains, and transport the majority of sediment to or along low-order stream channels (Jacobson et al., 1993; Guthrie and Evans, 2007). Mass movements can exert diverse influences on valley geometry and the spatial arrangement of channels in drainage networks (Korup et al., 2010; Korup, 2013). Hovius et al. (1998) describe three phases of valley development in Papua New Guinea: initial incision of isolated gorges, lateral expansion and branching by landsliding in patterns influenced by groundwater seepage, and entrenchment by river incision of landslide scars and deposits. Large mass movements can cause substantial sediment accumulation in valley floors, interrupting progressive channel incision into bedrock, as documented along the Indus River in Pakistan by Burbank et al. (1996). The downstream end of mass movement deposition in a valley bottom can create a knickpoint, or steeper segment, along a river's course. Landslide dams can pond river water upstream, but are typically breached within a few days (Costa and Schuster, 1988), resulting in an outburst flood with a peak discharge and sediment transport capacity that greatly exceeds precipitation-generated floods along the river (Cenderelli and Wohl, 2001, 2003). Small catchments can form in the detachment area of a mass movement. Large landslides can move so much mass that they reduce local elevation, cause drainage divides to shift location abruptly, or truncate headwater streams and cause stream piracy. Landslides can even cause drainage reversal. Mass movements also strongly influence spatial and temporal variations in sediment delivery to rivers in steep terrains (Korup et al., 2004, 2010). Mass movements can also occur in low-relief areas, particularly where an impermeable layer such as frozen ground or shale promotes saturation of overlying materials.

Creep occurs when particles displaced by bioturbation and in wetting–drying or freeze–thaw cycles move downslope under the influence of gravity (Figure S2.7) (Kirkby, 1967). Bioturbation via tree throw (Figure S2.8a) and burrowing by soil-dwelling rodents (Figure S2.8b), in particular, can displace substantial amounts of sediment (Heimsath et al., 1997; Roering et al., 2002; Gallaway et al., 2009). Creep in cold climates is strongly influenced by freeze–thaw processes (Hales and Roering, 2007) and, with increasing water or ice content, creep grades into solifluction or gelifluction (Figure S2.9), respectively, which involve the very slow downslope flow of partially saturated regolith. Creep is greatest in the upper meter of the soil and is proportional to surface gradient (Selby, 1982; McKean et al., 1993). This results in soil flux proportional to the depth-slope product (Furbish et al., 2009b).

Gradual diffusive processes in which individual grains move downslope typically occur via rainsplash or overland flow. *Rainsplash* occurs when rain falling on a surface loosens or detaches individual particles, making the particles more susceptible to entrainment by overland flow (Furbish et al., 2009a; Dunne et al., 2010) (Figure S2.10a). Overland flow may be capable of eroding measurable quantities of sediment where unvegetated, unfrozen slope surfaces are exposed (Dingwall, 1972; Rustomji and Prosser, 2001). *Thread flow* occurs when overland flow goes around individual roughness elements. Sediment can be stripped evenly from a slope crest and upper zone during *sheet flow* that submerges individual roughness elements and forms a fairly continuous sheet of water across the slope (Figure S2.10b). Erosion by sheet flow is particularly effective where inter-particle cohesion has been reduced by needle ice, trampling, or disturbance to vegetation (Selby, 1982). Microbiotic soil crusts, very coarse particles at the surface (Figure S2.10c), or vegetation cover can substantially reduce sediment detachment and erosion by rainsplash or sheet flow (Uchida et al., 2000). The presence of seasonally or permanently frozen soil can enhance overland flow and soil erosion during the melt season by impeding infiltration (Ollesch et al., 2006).

Horton (1945) proposed the existence of a *critical length of slope*, X_c, above which no erosion occurs via sheet flow. The critical length is a function of surface resistance and slope gradient. As a simple approximation, sheet flow erodes particles as a function of the shear stress τ exerted on the surface by the flowing water ($\tau = \gamma\ H\ S$, where γ is the specific weight of water, H is flow depth, and S is hillslope gradient) relative to the resistance of the surface over which the water is flowing. For a given surface resistance and flow depth, X_c will be shorter on steeper slopes up to about 40 degrees. Sheet flow becomes competent to transport sediment within a few meters of the drainage divide on long hillslopes subject to Hortonian overland flow, but microtopographic mounds (Figure S2.11) generated mostly by biotic processes can force the sheetwash to converge and diverge, increasing depth, velocity, and transport capacity in the converging zones. Sediment is released from microtopographic mounds into the sheet flow, however, so that the sediment supply is sufficient to prevent rill incision on the upper portion of the hillslope (Dunne et al., 1995).

At some point downslope, surface irregularities concentrate overland flow into slight depressions that then enlarge as the increasing water depth increases the shear stress acting on the substrate at the base of the flow. This can give rise to *rills* and *gullies*, typically described as parallel channels with few or no tributaries. Sometimes these names are used interchangeably, but rills can be distinguished as channels sufficiently small to be smoothed by ordinary farm tillage, whereas gullies are sufficiently deep not to be destroyed by ordinary tillage. Rills and gullies can form effective conduits for sediment erosion down slopes and into river networks (Sutherland, 1991), and rill-channel networks can exhibit equilibrium scaling characteristics for bifurcation and channel length ratios similar to those of river networks (Raff et al., 2004). Downslope movement of sediment tends to be extremely spatially and temporally variable as a result of local changes in slope gradient, ground cover, vegetation and microtopography (Saynor et al., 1994). Most hillslopes are shaped through time by some combination of multiple processes (Jimenez Sanchez, 2002). Just as subsurface water in a hillslope can be "old water" stored for relatively long periods of time between precipitation inputs, sediment that begins to move downslope and into a channel network can be stored for 10^3 years even in mountainous uplands. Investigating the Oregon Coast Range, for example, Lancaster and Casebeer (2007) found that sediment transit times along colluvial valley segments dominated by debris flows averaged 440 years, whereas the average transit time rose to 1220 years within the headwaters fluvial portion of the channel network. In both debris-flow and fluvial channels, significant volumes of sediment remain in storage for thousands of years. Consequently, terraces, fans, and some floodplain deposits near tributary junctions are effectively sediment reservoirs at time spans of 10^2–10^3 years (Lancaster et al., 2010).

Supplemental Section 2.2.2 includes more information on measuring and modeling the downslope pathways of sediment. Section 8.1.2 examines how human-induced changes in land cover influence downslope sediment movement by altering topography, vegetation, and soil depth and compaction.

2.2.3 Processes and patterns of water chemistry entering channels

The chemistry of precipitation falling over a drainage basin varies with distance from the ocean, with pollution inputs, and through time. The precipitation then reacts with plants, soil, regolith, and bedrock. The resulting chemistry of water entering a river is usually more influenced by the hillslope flow paths followed by the water than by the chemistry of the original precipitation. And, just as runoff during a storm can be dominated by old water that has been stored in the subsurface for some period prior to the storm, runoff chemistry can be dominated by old water (Anderson and Dietrich, 2001).

Waters entering a river via precipitation falling directly on the river, overland flow, soil water, and groundwater typically have distinctly different chemistries (McDonnell et al., 1991). The primary constituents of river chemistry (Berner and Berner, 1987; Allan, 1995) are

- the dissolved ions HCO_3^-, Ca^{2+}, SO_4^{2-}, H_4SiO_4, Cl^-, Na^+, Mg^{2+}, and K^+;
- dissolved nutrients N and P;
- dissolved organic matter;
- dissolved gases N_2, CO_2, and O_2; and
- trace metals.

The sum of the concentrations of the dissolved major ions is known as the *total dissolved solids* (TDS), and is typically highly temporally and spatially variable in response to factors such as precipitation input and downslope flow path, discharge, lithology, the growth cycles of terrestrial vegetation, and any factor that influences rates and processes of bedrock weathering and soil development (Berner and Berner, 1987). These factors include: topographic relief, climate, glaciation, snow cover, land use, and episodic events such as mass movements or volcanic eruptions. Supplemental Section 2.2.3 contains further information on controls on upland water chemistry, as well as a summary of measuring and modeling the chemistry of water entering channels.

Although water chemistry has traditionally been neglected by fluvial geomorphologists, understanding influences on the chemistry of water before the water enters the channel network and once the water is in the channel network is important for at least one reason: a significant proportion of management of river process and form is now undertaken in order to meet legislated water quality standards. River characteristics such as flow depth, hyporheic exchange, substrate grain-size distribution and stability, as well as the characteristics of aquatic and riparian communities, can strongly influence the chemistry of water in a river. These characteristics can potentially be managed to reduce or remove contaminants or other undesirable traits of water chemistry that result from the downslope pathways that water follows before entering a channel.

2.2.4 Influence of the riparian zone on fluxes into channels

Riparian is a Latin word designating something "of or belonging to the bank of a river" (Naiman et al., 2005). The *riparian zone* is the interface between terrestrial and aquatic ecosystems, and includes sharp gradients of environmental factors, ecological processes, and biotic communities (Gregory et al., 1991). Riparian zones can be difficult to delineate because they include components as diverse as depressions that create floodplain wetlands and higher-elevation natural levees (Figure 2.5).

The presence of a riparian zone can strongly influence the characteristics of water, sediment and solutes entering a river. Hillslope waters tend to be chemically and isotopically distinct from riparian zone waters and the degree of expression of hillslope

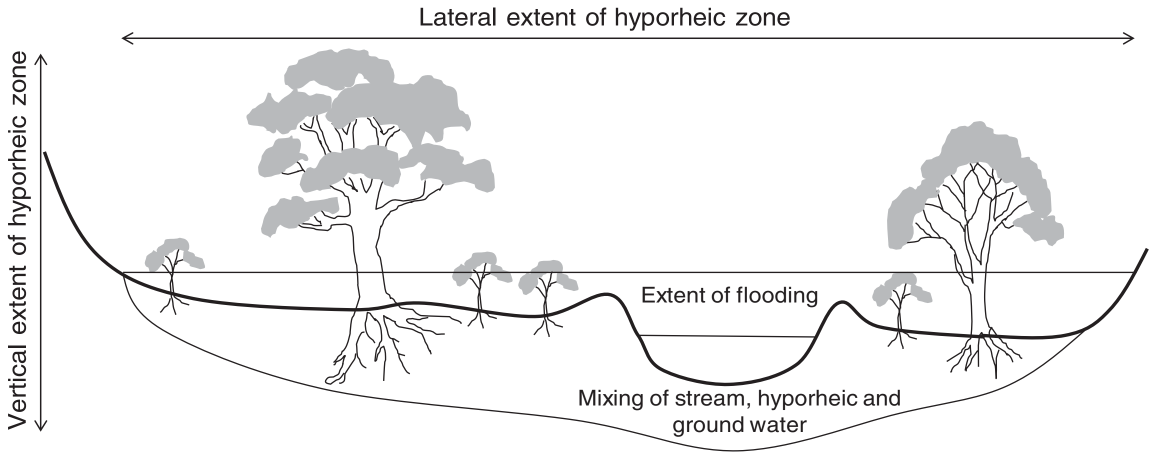

Figure 2.5 Schematic illustration of the spatial boundaries of a riparian zone. The boundaries of the riparian zone extend: outward to the limits of flooding—typically defined for a relatively frequent recurrence interval of flood, such as 10 years; upward into the canopy of streamside vegetation; and downward into the zone where groundwater, stream water, and hyporheic water mix.

water in the river can be minimal because of chemical reactions as the water passes through the riparian zone (Burns et al., 2001; McDonnell, 2003). Subsurface water coming from the riparian zone commonly leads the river hydrograph, with hillslope input either negligible or dominating the recession limb of the river hydrograph after the threshold for activating hillslope pathways is exceeded (McGlynn and McDonnell, 2003). Observations such as these led McDonnell (2003) to suggest that a catchment should be conceptualized as a series of reservoirs with coupled unsaturated and saturated zones, and explicit dimensions and porosities, which connect laterally and vertically through time and space in linear and nonlinear ways.

Vegetated riparian zones can effectively trap and store sediment of sand size or finer that is being transported by overland flow or by overbank flows from a channel (Naiman et al., 2005). Sediment deposition reflects the greater hydraulic roughness, and consequently lower velocities and transport capacities, of flow passing over and through the stems and leaves of the vegetation (Hickin, 1984; Griffin et al., 2005). Once the sediment is deposited, plant roots facilitate stabilization and continued storage of the sediment (Allmendinger et al., 2005; Tal and Paola, 2007). Many riparian species are able to grow successive sets of near-surface roots as the plant's base is progressively buried by sedimentation (Figure S2.12).

Numerous studies indicate that riparian zones can serve as sinks or buffers for nitrates in agricultural watersheds (Mitsch et al., 2001). The capacity of the riparian zone to retain dissolved and particulate nutrients such as nitrogen, phosphorus, calcium, and magnesium is controlled by two factors. The first is the hydrologic characteristics of the riparian environment, including water table depth, water residence time, and degree of contact between soil and groundwater. The second level of control involves biotic processes, including plant uptake and denitrification. Denitrification is microbially facilitated nitrate reduction that ultimately produces molecular nitrogen, N_2 (Naiman et al., 2005). As an important nutrient and component of greenhouse gases, nitrogen has received particular attention, and the complexities of nitrogen dynamics illustrate the importance of riparian hydrology and biology (Supplemental Section 2.2.4).

Biogeochemical hot spots for dissolved nutrients include anoxic zones beneath riparian environments because of the microbial communities present in these zones (Lowrance et al., 1984). Riparian communities in wet temperate and tropical environments can mediate nitrate fluxes into rivers because denitrification and plant uptake remove allochthonous nitrate (nitrate that originated elsewhere, such as from groundwater) within a few meters of travel along shallow riparian flow paths (McClain et al., 1999). The effectiveness of this nitrogen removal can vary seasonally from very high values in late summer to much lower values in winter in temperate regions (Maitre et al., 2003). In contrast, little nitrogen processing may occur during transport from the uplands through riparian areas in arid regions if precipitation moves rapidly across the riparian zone as surface runoff (McClain et al., 1999).

In addition to uptake of nutrients by vascular plants and microbial communities, alluvial surfaces of different ages within the riparian zone can have different soil and water chemistry and infiltration because vegetation influences evaporation and transpiration that control soil water potential and concentration of salts. Vegetation also creates litter, a surface layer of dead plant parts such as leaves or conifer needles (Figure S2.13) that influences infiltration and evaporation from the soil, as well as contributing organic matter to the soil. Some plants also directly fix nitrogen in the soil (Van Cleve et al., 1993).

People have substantially reduced the extent of riparian vegetation and floodplain wetlands worldwide, causing associated changes in fluxes of water, sediment, and dissolved and particulate materials from these environments into channels. Although no global estimates of total loss or degradation of riparian ecosystems have been published, numerous regional or basin-wide estimates from industrialized countries suggest losses of much greater than half of the riparian zones along rivers in industrialized countries, with consequent loss of flood attenuation, sediment storage, and biological processing of nutrients.

2.3 Channel initiation

Channel initiation is a threshold phenomenon in which surface or subsurface flow concentrates and persists sufficiently to produce a discrete channel. The upstream boundary of concentrated water flow and sediment transport between definable banks is the *channel head* (Figure S2.14) (Montgomery and Dietrich, 1988, 1989), which separates the process domains of hillslopes or unchanneled hollows (Figure S2.15) from channel networks (Dietrich and Dunne, 1993). Banks can be defined in the field based on the presence of sediment transport (wash marks, small bedforms, armored surfaces) and an observable sharp break in slope. The channel head does not necessarily coincide with the location where perennial flow occurs, which is the *stream head* (Figure S2.16). Channel segments of ephemeral and intermittent flow can be present above the stream head and below the channel head even in wet regions.

The locations of individual channel heads in even a small channel network can have substantially different drainage areas. This is not surprising, given the multiple factors that influence the location of a channel head. These factors include gradient, drainage area, infiltration capacity, porosity and permeability, and cohesion, each of which influences surface and subsurface flow paths and erodibility of near-surface materials (Figure S2.17). The location of a channel head can also vary through time as a result of changes in climate or land cover that affect runoff, surface erodibility, and sediment supply (Montgomery and Dietrich, 1992). Following a wildfire that killed vegetation and burned the surface layer of litter and duff on forested hillslopes in the semiarid Colorado Front Range, USA, surface runoff became more common, causing channel heads to migrate upslope and form at minimum drainage areas two orders of magnitude smaller than pre-fire minimum drainage areas (Wohl, 2013a) (Figure S2.18).

The distribution of channel heads across a drainage basin can reflect primarily surface or subsurface flow, some combination of the two, or mass movements (Figure 2.6). Regardless of the mechanisms, flow convergence facilitated by topography and/or stratigraphy promotes the concentration of flow that initiates channels (Dunne, 1990; Dietrich and Dunne, 1993).

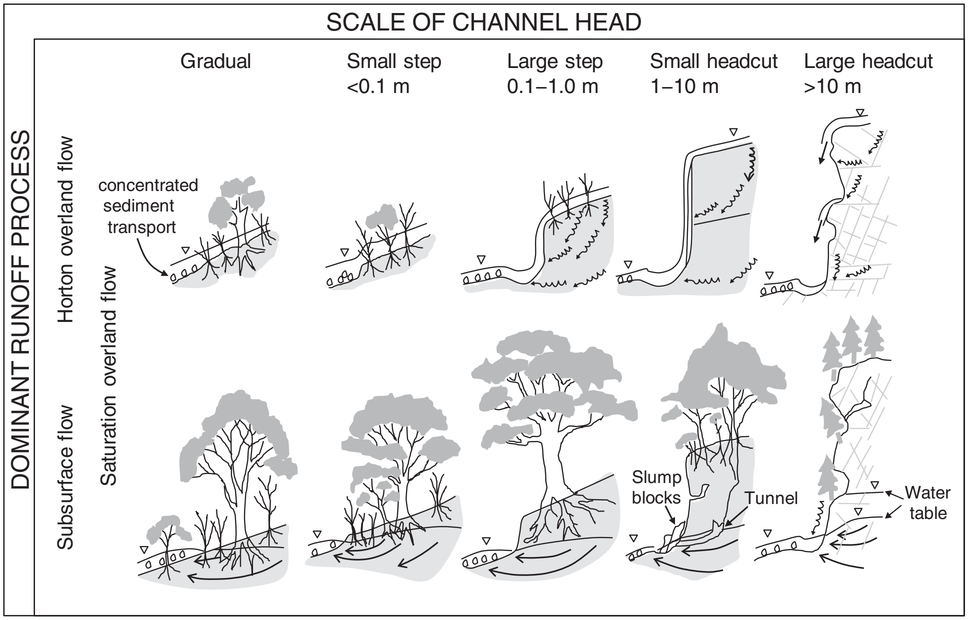

Figure 2.6 Classification of channel heads based on incision depth and dominant runoff process. Sketches indicate flow paths for Hortonian overland flow and subsurface flow. Smooth arrows indicate saturated flow; wiggly arrows indicate unsaturated percolation, including flow through macropores. Even at sites with substantial Hortonian overland flow, the face of a large headcut can allow the emergence of erosive seepage. Saturation overland flow drives erosion that includes features from both of the other runoff types (from Dietrich and Dunne, 1993, Figure 7.6).

Channel heads that reflect primarily surface processes can form via Hortonian or saturation overland flow that leads to rilling. The greatest surface irregularities and erodibility result in flow concentrated in a *master rill* to which adjacent slopes are cross-graded (Horton, 1945). Steep slopes, high rainfall intensity, low infiltration rates, and high erodibility favor rilling (Dietrich and Dunne, 1993). These conditions are characteristic of arid, semiarid, or disturbed landscapes dominated by Hortonian overland flow (Kampf and Mirus, 2013).

Rills can develop nearly simultaneously across a terrain and then integrate into a network (Dunne, 1980), although drylands typically have discontinuous headwater networks of short, actively eroding channel reaches separated by unchanneled or weakly channeled, vegetated, stable reaches (Tucker et al., 2006). Alternatively, channels can extend downstream during slow warping or intermittent exposure of new land on a rising land surface, or channels can extend upstream in response to an increase in slope or the lowering of base level (Dunne, 1980).

Erosional hot spots occur where topographic constrictions or locally steep gradients amplify hydraulic forces sufficiently to overcome surface resistance and initiate *headcuts* (Tucker et al., 2006), which may coincide with the channel head or occur downstream. Headcuts that coincide with the channel head are vertical faces that separate upslope unchanneled environments from downslope channels. Headcuts downstream from the channel head separate upslope presently stable channel segments from downslope recently incised channel segments (Figure 2.7).

Channel heads dominated by subsurface processes can form via piping or sapping, or from shallow landsliding on steep slopes that creates a topographic low where subsurface flow can begin to exfiltrate (Montgomery et al., 2002; Kampf and Mirus, 2013). *Piping* occurs in the unsaturated zone when flow is sufficiently concentrated to erode or

Figure 2.7 View upstream to a headcut along an ephemeral channel in the grasslands of eastern Colorado, USA. The headcut is just over 2 m tall. Approximately 1 m to the left of the survey rod and circled is a buried tree stump indicating a past cycle of incision followed by aggradation, prior to the present phase of incision. The channel upstream of this headcut is stable now, but has incised in the past.

dissolve subsurface materials and create physical conduits for preferential flow. If a pipe enlarges sufficiently to cause collapse of overlying materials, the pipe can have a surface expression as a channel head (Figure S2.19). Piping typically occurs in unconsolidated material.

Sapping occurs in the saturated zone, which can intersect the surface to form a spring. The return of subsurface flow to the surface enhances physical and chemical weathering and creates a pore-pressure gradient that exerts a drag on the weathered material (Dunne, 1980). Sapping can occur in unconsolidated material or bedrock, but is not likely to be an effective mechanism of erosion in bedrock (Lamb et al., 2006; Jefferson et al., 2010) unless it is accompanied by chemical dissolution in carbonate rocks.

Early field-based studies of channel initiation indicated that the source area above the channel head decreases with increasing local valley gradient in steep, humid landscapes with soil cover (Montgomery and Dietrich, 1988). For slopes of equal gradient, source area can vary in relation to total precipitation or precipitation intensity, as these characteristics influence concentration of runoff (Henkle et al., 2011). Empirical, site-specific relations between topographic parameters typically use bounding equations to quantify the range in channel head locations (Montgomery and Dietrich, 1989, 1992). An example comes from the work by Prosser and Abernethy (1996) in Gungoandra Creek, Australia

$$A = 30 \tan \theta^{-16} \tag{2.1}$$

In this case, A is specific catchment area (the ratio of upslope catchment area to lower contour width), and $\tan\theta$ is hillslope gradient.

The inverse slope–area relationship is not always present, however, because of differences in runoff processes. Low-gradient hollows with convergent topography and seepage erosion can differ from steeper topography where channel initiation is more likely to reflect Hortonian or saturation overland flow or even landsliding (Montgomery and Dietrich, 1989; Montgomery and Foufoula-Georgiou, 1993). In terrains with substantial flow through fractured bedrock, bedrock topography is likely to exert a greater influence on channel head locations than does surface topography (McDonnell, 2003; Adams and Spotila, 2005; Jaeger et al., 2007). Heterogeneities in the bedrock, such as spatial variation in joint density, can influence both the location of channel heads and the spatial distribution of channels within a river network, with channels tending to follow more densely jointed bedrock (Loye et al., 2012).

If channel heads form where saturation overland flow exerts a boundary shear stress that exceeds the critical value for substrate erosion, the channel initiation threshold, C, can be expressed as the product of contributing catchment area, A, and hillslope gradient, S (Dietrich et al., 1992, 1993)

$$AS^a \geq C \tag{2.2}$$

Substantial variability in values of A and S reflects the influence of factors such as vegetation, slope aspect, surface versus subsurface flow paths, and substrate grain size (Montgomery and Foufoula-Georgiou, 1993; Prosser et al., 1995; Istanbulluoglu

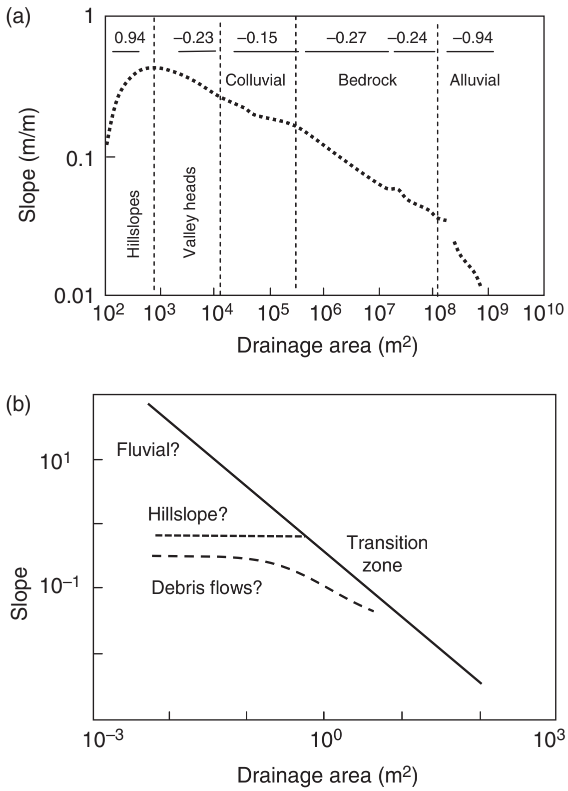

Figure 2.8 (a) Plot of log-bin averaged drainage area versus slope relationship for the Olympic Mountains of Washington, USA. Plot shows the mean slope of individual 10 m grid cells for each 0.1 log interval in drainage area. Numbers at the top are the exponents for a power function regression of values in the segments of the plot indicated by horizontal lines below. Dashed vertical lines divide the plot into areas considered to reflect different geomorphic zones of the landscape, or process domains (from Montgomery, 2001, Figure 5A). (b) Hypothetical topographic signatures for hillslope and valley processes. Area and slope are measured incrementally up valley mainstem to the valley head (from Stock and Dietrich, 2003, Figure 1a; Wohl, 2010, Figure 2.4).

et al., 2002; Yetemen et al., 2010) (Figure 2.8). In the absence of field-based data, many investigators assume that channel heads lie near reversals or inflections in averaged hillslope profiles (Ijjasz-Vasquez and Bras, 1995), although this can result in significant over- or under-estimations of contributing area (Tarolli and Dalla Fontana, 2009; Henkle et al., 2011) (Figure S2.20).

Human-induced changes in land cover can alter the location of channel and stream heads by changing infiltration and the balance of water conveyed downslope in surface and subsurface flow paths. The most typical scenario involves decreased infiltration and smaller contributing areas for channel and stream heads (e.g., Montgomery, 1994).

2.4 Extension and development of the drainage network

Glock (1931) proposed that networks go through five stages of development. Channel heads form and rills integrate through cross-grading during the stage of initiation. The stage of extension by elongation features headward growth of channels. During stage three, extension by elaboration, tributaries are added. This results in a stage of maximum extension, followed by the final stage of abstraction as local relief is reduced and tributaries are lost (Figure 2.9). Subsequent field-based studies that substitute surfaces of different ages for time (e.g., successive basalt flows or glacial deposits in the same region) tend to support this general model of network development, as do physical experiments (Schumm et al., 1987).

Supplemental Section 2.4 describes different types of numerical models used to study network development through time.

2.4.1 Morphometric indices and scaling laws

Initial descriptions of drainage networks focused on their relationship to topography. An example is J.W. Powell's nineteenth century characterization of antecedent and superimposed drainage networks. Powell inferred that a river cuts across a mountain range rather than flowing down and away from both sides of a mountain drainage divide because the river had maintained its location and cut downward as the surface was deformed around it (*antecedent*), or the

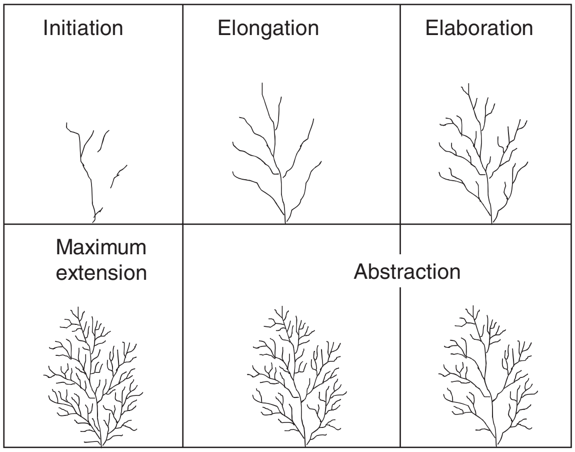

Figure 2.9 Illustration of Glock's stages of drainage network development. During initiation, a few, potentially longitudinally discontinuous channels are present. With elongation, these channels grow upslope and integrate into a continuous network. Additional, shorter, first-order streams are added during elaboration and maximum extension. As secondary slopes and small-scale relief are reduced by continued channel erosion, some first-order streams are lost along the downstream margins of the main channel during abstraction.

river had eroded down to a buried structure (*superimposed*). Early descriptions also emphasized the categorization of networks based on their planform appearance (Zernitz, 1932; Parvis, 1950). Networks with a dendritic appearance typically indicate relatively low relief, homogeneous substrate, for example, whereas a rectangular network could indicate strong regional joint control that facilitates right-angle channel junctions where joints intersect.

Starting in the mid-twentieth century with work by Horton (1932, 1945), descriptions of drainage networks became increasingly quantitative and nondimensional. Horton, Schumm (1956), and Strahler (1957) modified earlier morphometric indices and developed new indices, helping to disseminate the idea of characterizing river networks using stream order, bifurcation ratio, length ratio, drainage density, relief ratio, and other indices.

Stream order is a number assigned to a stream segment. A segment with no tributaries is a first-order stream. Where two first-order segments join, they form a second-order segment; two second-order segments form a third-order segment, and so on (Strahler, 1952). *Bifurcation ratio* is the ratio of the number of streams of a given order to the number of streams of the next higher order. *Length ratio* is ratio of the average length of streams of a given order to the average length of streams of the next higher order. *Drainage density* is the ratio of total length of streams to basin area. *Relief ratio* is the ratio of total basin relief to basin length.

The various morphometric indices were designed to facilitate comparisons across channel networks of diverse sizes, and to provide insight into underlying, fundamental characteristics of energy distribution, erosion, and the distribution of channels across a landscape. Horton (1945), for example, proposed a law of stream numbers and a law of stream lengths and interpreted their consistency across networks as suggesting independence of the specific geomorphic processes in any particular network. Hack (1957) proposed that consistent scaling relations could be used to describe network characteristics such as the relation of stream length L to drainage area A

$$L = 1.4\,A^{0.6} \tag{2.3}$$

Shreve (1966) and Smart (1968) compared statistical properties among populations of natural channels not strongly influenced by geologic controls. They interpreted statistical similarity as indication that the networks were topologically random and thus a consequence of random development according to the laws of chance. Milton (1966) and Kirchner (1993), however, proposed that, as all branching phenomena could equally satisfy the Horton criteria, the laws of drainage network configuration were geomorphologically irrelevant.

Similar controversy has arisen over more recent work describing drainage network patterns using fractal approaches (e.g., Rodríguez-Iturbe and Rinaldo, 1997). *Fractals* are geometrical structures with irregular shapes that retain the same degree of irregularity at all scales and are thus self-similar. Investigators who have observed fractal properties in networks argue that this reflects a scale independence to landscape dissection. Beauvais and Montgomery (1997), however, note that a minimum contributing area is needed to concentrate surface or subsurface flow sufficiently to form channels, indicating a scale dependence to landscape dissection. Beauvais and Montgomery found that drainage networks in the western United States do not exhibit the scaling properties required for fractal geometry.

Regardless of the insights that may or may not result from statistical self-similarity or consistent power functions between discrete variables, individual morphometric indices remain useful for inter-drainage comparisons and for understanding drainage network history. Stream order, drainage density, and relief ratio are especially widely used (Figure 2.10).

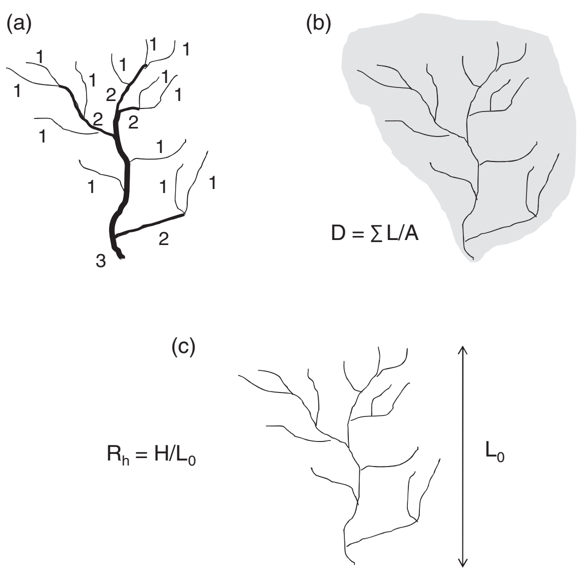

Figure 2.10 Examples of commonly used stream morphometric indices. (a) Stream order from Strahler (1952). First-order streams have no tributaries. A higher stream order occurs where two streams of equal order join. This stream network is third order. Each of the first-order streams is also an exterior link, and each of the second- and third-order streams is an interior link (Shreve, 1966). (b) Drainage density is the ratio of the sum of stream lengths per unit drainage area (area shaded gray). (c) Relief ratio R_h is the ratio of the maximum topographic relief (elevation difference) H to the longest horizontal distance of the basin measured parallel to the main stream L_0.

Stream order provides a convenient means to indicate the relative size of a channel segment (e.g., a first-order stream) or of a drainage network (e.g., a third-order drainage). Stream order is sometimes replaced by a link as the basic unit of network composition. A *link* is an unbroken section of channel between successive nodes (sources, junction, or outlet) (Shreve, 1966). An *exterior link* corresponds to a first-order stream in that it extends from a source to the first junction downstream. An *interior link* connects two successive junctions, or the last junction and the outlet (Knighton, 1998).

Drainage density reflects the degree to which a basin is dissected by channelized flow, and this in turn reflects the relative influences of substrate resistance (higher resistance typically results in lower density), rainfall-runoff-land cover (high drainage densities typically occur in semiarid regions that receive more rainfall than arid regions, but lack the continuous vegetation cover and associated higher infiltration of humid regions), and age of the network (as reflected in Glock's stages of drainage development).

Relief ratio indicates the steepness of topography within a basin. Steepness can provide insight into tectonic uplift and rock resistance, as well as downslope pathways and storage time of water, sediment, and solutes.

A fundamental limitation on being able to characterize any drainage network is the ability to accurately map actual channel locations. This is not particularly difficult for large channels, but can be very problematic for the smallest, headwater tributaries. The default assumption is typically to use the extent of "blue lines" used to indicate channels on topographic maps or digital elevation models (DEMs). The correspondence between these lines and real channels varies with the spatial resolution and type of data used to generate the map or DEM, and the consistency of channel extent through time (e.g., channel extent can vary substantially over a period of years in drylands or in areas with rapidly eroding headwaters). Another approach is to use contour crenulations (the degree to which contour lines are distorted when crossing a valley) as indicators of sufficient valley incision to create channelized flow, but the threshold value chosen for crenulation varies between studies and the correspondence between crenulation and channelized flow varies between regions. As noted in Section 2.3, reversals or inflections in averaged hillslope profiles can also be used to indicate the location at which channelized flow begins across a drainage basin. Each of these methods is subject to errors, and it is important to be aware of these errors as a source of uncertainty in network analysis.

2.4.2 Optimality

Various aspects of river form and process have been described using *extremal hypotheses* that characterize tendencies toward which rivers evolve based on the balance between energy available to move water and sediment and shape channels, and the resistance of the channel boundaries. The extremal hypothesis of *optimal channel networks* has been applied to understanding the spatial arrangement of channels in a drainage network. Optimal channel networks display three principles of optimal energy expenditure: (1) minimum energy expenditure in any link of the network; (2) equal energy expenditure per unit area of channel anywhere in the network; and (3) minimum total energy expenditure in the network as a whole (Rodríguez-Iturbe et al., 1992). The effect of the tendencies underlying these principles can be considered in terms of local and global optimal energy expenditure. In *local optimal energy expenditure*, the channel properties (e.g., channel cross-sectional geometry and gradient) of river networks adjust toward a constant rate of energy dissipation per unit channel area. In *global optimal energy expenditure*, the topological structure of networks (e.g., drainage density, bifurcation ratio) adjusts to minimize total energy dissipation rate (Molnár and Ramírez, 1998; Molnár, 2013). Energy expenditure can be quantified using variables such as shear stress or stream power.

These principles largely grew out of work by Leopold and Langbein (1962), Langbein (1964), and Langbein and Leopold (1964). Leopold and Langbein proposed that the mean form of river channels represents a *quasi-equilibrium state* that is most probable to occur because it balances the opposing tendencies of minimum total rate of work in the whole river system and uniform distribution of energy expenditure throughout the system.

The tendency toward uniform distribution of energy expenditure can be understood by considering a portion of channel with higher rates of energy expended against the channel boundaries because of a lateral constriction or larger gradient that results in locally higher velocity. Assuming that the channel boundary is erodible, the greater rate of energy directed against the boundary should enlarge the boundary until the rate of energy expenditure declines. An example comes from the Colorado River in the Grand Canyon, USA. Flash floods and debris flows along steep, ephemeral tributaries to the Colorado River create tributary debris fans that constrict the main channel. During large floods on the Colorado River, these constrictions force the flow to become critical or supercritical (Section 3.3.1). The higher energy directed against the channel boundaries through the constriction during floods erodes the toe of each debris fan until the main channel is sufficiently widened to permit subcritical flow even during large floods (Kieffer, 1989). Erosion then decreases and the channel cross-sectional geometry becomes stable until the next tributary input again creates a constriction. In other words, sites of very high energy expenditure are transient in an erodible channel. Channel geometry will adjust to reduce energy expenditure at the site and create more uniform distribution of energy expenditure relative to upstream and downstream sites. Numerous adjustments of the type described for the Grand Canyon along the course of a river or throughout a drainage network can result in uniformity and minimization of energy expenditure, given sufficient time and energy relative to boundary resistance.

One of the points of discussion regarding extremal hypotheses is whether most natural channels ever actually attain an optimal state, given the potential for highly resistant boundaries or continual perturbations such as tectonic uplift or sediment inputs. An assumption underlying hypotheses of optimal channel networks is that place-specific details such as lithology, structure, or tectonics that influence substrate resistance are absent or negligible. This is often not the case. Consequently, extremal hypotheses may describe the conditions toward which a system is evolving, rather than the actual state of the system. Many natural river networks, however, do approximate the structure of optimal networks (Rodríguez-Iturbe and Rinaldo, 1997; Molnár and Ramírez, 1998).

2.5 Spatial differentiation within drainage basins

Schumm (1977) conceptualized a drainage basin as being longitudinally zoned with respect to sediment dynamics. He distinguished a headwater production zone, mid-basin transfer zone, and downstream depositional zone (Figure S2.21). These zones represent predominant processes: the headwater zone does include transfer and deposition, for example, but is most characterized by sediment production. Subsequent research has elaborated on and quantified these distinctions. Benda et al. (2005), for example, describe headwater streams as sediment reservoirs at timescales of 10^1–10^2 years, along which sediment is episodically evacuated by debris flows or gully erosion. Sediment yield per unit area, or *sediment delivery ratio*, decreases and sediment residence time increases as stream order and drainage area increase as a result of increasing storage on hillslopes or valley bottoms (Schumm and Hadley, 1961; Dietrich and Dunne, 1978). These changes may also be a step function, with relatively large rates of change in sediment storage volume and time at transitions in geomorphic processes, such as where debris flows give way to predominantly fluvial processes (Lancaster and Casebeer, 2007).

Rice (1994) described a continuum of sediment transfer to channels, which represents another way of describing the sediment connectivity and storage discussed in Chapter 1. At one end of the continuum are *strongly coupled links* in which sediment is transferred from hillslopes to channels relatively rapidly and continuously. At the other end of the continuum are *completely buffered links* along which floodplains or valley-fill deposits protect hillslopes from basal erosion and limit direct sediment supply from hillslopes to channels. Schumm's headwater production zone would be best characterized as composed of strongly coupled links, whereas the proportion of completely buffered links would increase progressively downstream in the transfer and depositional zones. Again, these are general patterns: headwater channels can be buffered and downstream channel segments can be strongly coupled. Most importantly, the degree of coupling for any channel segment can vary through time.

Spatial differentiation of geomorphic process domains can be used to distinguish six basic process domains (Montgomery et al., 1996; Sklar and Dietrich, 1998; Montgomery, 1999) (Figure 2.11).

- Hillslopes: *Hillslope* in this context refers to the unchannelized, largely straight or convex portion of slopes, typically dominated by diffusive transport or slopewash.
- Unchanneled hollows or zero-order basins: *Unchanneled hollows* are hillslope concavities that do not have channelized flow, but serve as sediment storage sites and are important points for the initiation of mass movements (Dietrich and Dunne, 1978; Montgomery et al., 2009).
- Debris-flow channels: *Debris-flow channels* can be influenced by fluvial processes, but non-fluvial processes dominate sediment dynamics and channel geometry.
- Bedrock-fluvial channels: *Bedrock-fluvial channels* have bedrock exposed along the channel boundaries or at sufficiently shallow depth that overlying alluvium is readily mobilized during higher flows, so that the underlying bedrock limits channel boundary erosion. The distinctions among bedrock-fluvial channels and coarse- and fine-bed alluvial channels reflect differences in the balance between sediment supply and river transport capacity. Where transport capacity exceeds sediment supply, bedrock is exposed in the channel bed.
- Coarse-bed alluvial channels: *Coarse-bed alluvial channels* are channels with unconsolidated

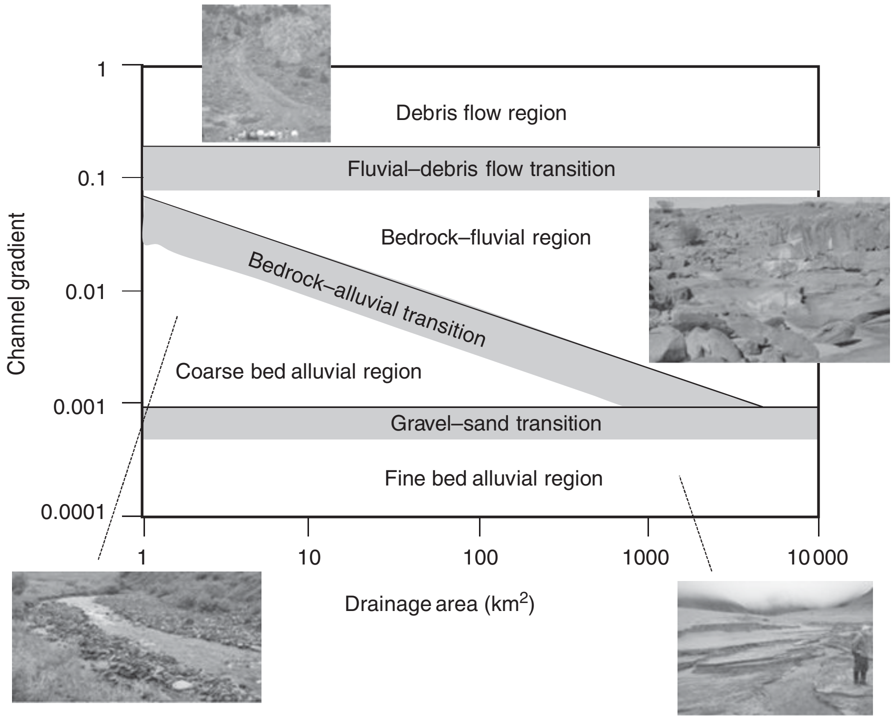

Figure 2.11 Hypothesized distribution of channelized erosional process domains (fluvial vs. debris flow) and channel substrate types in relation to drainage area and slope. (From Sklar and Dietrich, 1998, Figure 1, p. 241.)

sediment coarser than sand size. Sediment mobility in these channels is likely to be limited by sediment supply, rather than flow energy.

- Fine-bed alluvial channels: *Fine-bed alluvial channels* have non-cohesive sand-sized sediment forming the channel bed. Sediment mobility is more likely to be limited by flow energy than by sediment supply.

The transition between debris-flow and completely fluvial channels is marked by changes in channel process and form as a result of the differences in flow mechanics. Flow conditions can vary among debris flow, hyperconcentrated flow, and water flow downstream and with time during a flow as a result of changes in sediment concentration (O'Connor et al., 2001). Water and sediment move together in a *debris flow* as a single viscoplastic body that can be up to 90% sediment by weight, with a bulk density of 1.8–2.6 g/cm^3 (Costa, 1984; Iverson, 2005). A *hyperconcentrated flow* is a water flow with 40%–70% sediment by weight and a bulk density of 1.3–1.8 g/cm^3 (Costa, 1984; Pierson, 2005). A *water flow* carries only 1%–40% sediment by weight and has a bulk density in the range of 1–1.3 g/cm^3 (Costa, 1984).

Differences in bulk density are important because bedload transport rate and maximum clast size can increase with increasing fluid density if the flow around the grains is not laminar (Rickenmann, 1991). Debris flows can be highly erosive in steep or confined channels, and can create substantial aggradation in lower gradient, less confined channel segments or at channel junctions (Benda, 1990; Wohl and Pearthree, 1991). Because they are capable of mobilizing clast sizes and volumes of sediment greater than those mobilized by fluvial processes, debris flows drive cycles of aggradation and degradation in the mountainous portions of many river networks (Benda and Dunne, 1987), as well as strongly influencing the gradient of side slopes, disturbance regime, and valley and channel morphology, including aquatic habitat, organic matter, and the structure and composition of riparian vegetation (Swanson et al., 1987; Florsheim et al., 1991; Hewitt, 1998; Benda et al., 2003b).

Valleys consistently subject to different flow processes can have distinctly different geometries (De Scally et al., 2001). Valleys dominated by debris flows and fluvial erosion are more incised and contain closely spaced channels, for example, relative to valleys dominated by landslides. Valley gradient can decrease abruptly at the transition from debris flow to fluvial channels (Stock and Dietrich, 2006). Downstream from this transition strath terraces (Section 7.2.1) can form and drainage area–stream gradient relations (*A*–*S*) follow fluvial power laws such that *S* varies as an inverse power law of *A* (Stock and Dietrich, 2003). Debris flows in forested regions can entrain substantial volumes of wood that the debris flow subsequently deposits as a dam that can give rise to aggradation, terraces, and outburst floods (Lancaster et al., 2003; Comiti et al., 2008; Rigon et al., 2008). Mountainous headwater channels subject to debris flows can display downstream coarsening of median bed surface grain size (Brummer and Montgomery, 2003), in contrast to the more typical downstream fining in fluvial channels (Section 5.2.2).

2.6 Summary

Every process discussed in this chapter is characterized by diverse levels of spatial and temporal variability. Beginning with inputs and downslope movement of water, the details of how precipitation is produced and how water moves downslope via surface and subsurface pathways vary significantly across even a small catchment and through relatively brief intervals (hours to days) of time. Water movement, in particular, is characterized by thresholds. Once precipitation or glacier or snow melt begins, the downslope pathways and rates of water movement change substantially as different components of a hillslope turn on, such as when a subsurface pipe network becomes active or infiltration gives way to surface flow. Similarly, sediment produced through bedrock weathering does not move uniformly downslope into rivers, but instead moves abruptly during mass movements or diffusive creep that occurs primarily during precipitation, with sediment coming disproportionately from steep or otherwise more readily erodible portions of a catchment. Solutes entering the river come disproportionately from more readily weathered minerals or lithologies, and are strongly influenced by movement through biochemically active portions of the catchment, such as riparian zones.

Channel and stream heads that mark the start of a river network reflect the processes that govern downslope movement of water and sediment, as these processes cause water to concentrate sufficiently to create definable, persistent channels. Threshold conditions of drainage area and hillslope gradient at which channels form vary among different catchments and within a catchment through time in response to differences in surface and subsurface downslope pathways of water and sediment.

A variety of morphometric indices have been used to characterize the spatial distribution of channel networks and topography within catchments as a means of facilitating comparison among catchments. Stream order, drainage density, and relief ratio are particularly widely used. Investigators have also used mathematical frameworks such as optimality and fractals to search for universal physical laws that underlie and can provide insight into observed patterns in river networks. This approach remains controversial at least in part because it is unclear to what degree actual river networks, with their departures from ideal mathematical forms as a result of spatial variation in substrate resistance, sediment inputs, and so forth, can be effectively described by optimality or fractals. At a more general level, most catchments can be usefully viewed as being spatially differentiated into:

- headwaters that are characterized by relatively efficient movement of water, solutes, and sediment from uplands into first- and second-order channels;

- mid-basin zones characterized not only by downstream transfer of water, solutes, and sediment, but also by some storage of these materials in valley-bottom environments such as alluvial fans and floodplains; and
- lower basin depositional zones characterized by more spatially extensive and longer-term storage of water, solutes, and sediment in well-developed floodplains and deltas.

Channel processes I

Chapter 3
Water dynamics

This chapter introduces the basic physical properties of water flow within a channel. The discussion of hydraulics starts by explaining how flow is classified based on velocity, the ratio of viscous to inertial forces, and the ratio of inertial to gravity forces. The next section describes measures of the energy of water flowing in a channel and the implications of energy level for stability of the flow and adjustments of the channel boundaries. Energy and stability are closely connected to sources of flow resistance and equations used to quantify resistance, as well as velocity and turbulence that result from the interactions between available flow energy and resistance. The final section on hydraulics returns to the idea of energy and introduces variables used to express the energy exerted against the channel boundaries. Understanding the physical properties of flowing water is necessary to everything that follows in this volume because the quantity of energy available and the resistance created by the channel boundaries govern the movement of sediment and adjustment of channel geometry.

The second part of the chapter first addresses methods used to estimate volume of flow over differing time periods. This is followed by a discussion of the effects on stream discharge of surface–subsurface water exchanges, and the effects on discharge of flow regulation and channel engineering.

3.1 Hydraulics

Water flowing down a channel converts potential energy to kinetic energy and dissipates energy. The rate and manner in which energy are expended depend on the configuration of the channel, including the frictional resistance of the channel boundaries, and the amount of sediment being transported. Velocity is one of the most commonly measured hydraulic variables, and is particularly variable in time and space because of its sensitivity to frictional resistance. The basic flow continuity equation introduced in Chapter 1, $Q = w\, d\, v$, is quite mathematically simple. Any of the dependent variables can be very difficult to predict in natural channels, however, even if Q is known, because each is influenced by other properties of the flow and channel boundaries. Width, for example, reflects Q, and also the ability of flow to erode the channel banks. Depth reflects Q, as well as the erodibility of the bed. Velocity reflects Q, and also frictional resistance of the channel boundaries. Width, depth, and velocity also influence each other. Increasing flow depth changes the resistance to flow and hence the velocity associated with a particular bed grain-size distribution, for example, as the grains protrude into a progressively smaller proportion of the total flow depth.

Rivers in the Landscape: Science and Management, First Edition. Ellen Wohl.

Companion Website: www.wiley.com/go/wohl/riverslandscape

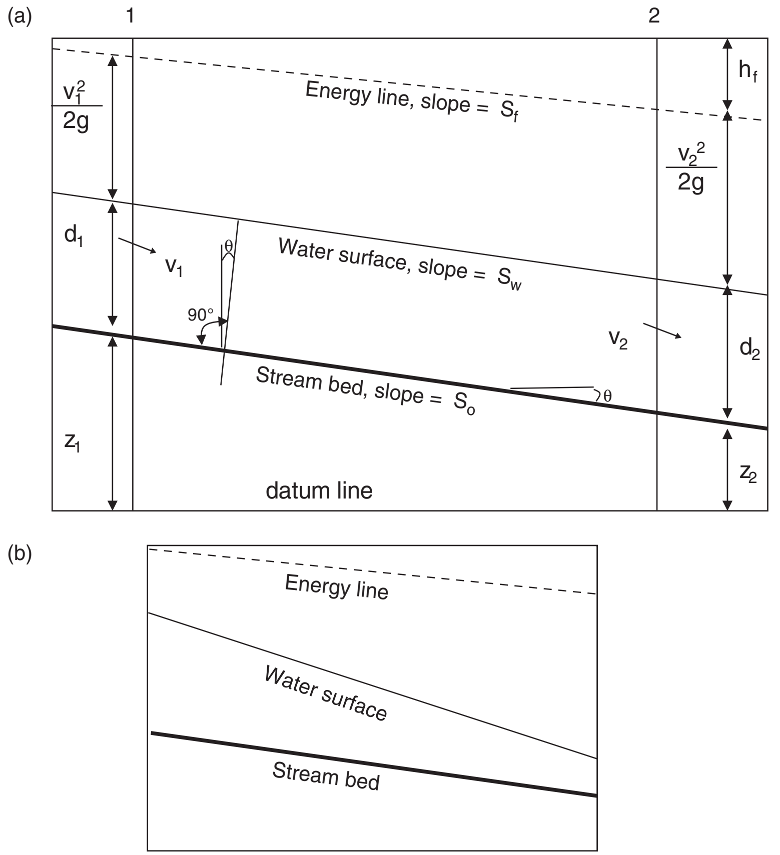

Figure 3.1 Parameters of open-channel flow; variables are defined in the text. Longitudinal view of a short channel segment, with flow from left to right. (a) Uniform flow; assumed parallel flow with a uniform velocity distribution and small channel slope. (b) Gradually varied flow, with velocity increasing and depth decreasing downstream.

Knowledge of basic hydraulic properties is necessary to understand the complex interactions among flowing water, channel geometry, and sediment transport, and to understand the assumptions that underlie many of the equations applied to processes in natural channels. Consequently, the first portion of this chapter introduces hydraulic parameters and equations commonly applied to natural channels.

3.1.1 Flow classification

Much of the following discussion of basic hydraulics in open-channel flow comes from Chow (1959), a classic hydraulics textbook, from Fox and McDonald (1978), a classic fluid mechanics textbook, and from Robert (2003), which contains a thorough and comprehensive treatment of water and sediment dynamics in rivers.

Water flowing in a conduit can be either open-channel flow or pipe flow. *Open-channel flow* has a free surface at the boundary between the water and the atmosphere, and this free surface is subject to atmospheric pressure (Chow, 1959). Figure 3.1 illustrates several important parameters used to characterize open-channel flow. For simplicity, individual flow lines are parallel and moving at the same velocity, and the slope of the channel is small. The total energy in the flow of the section with reference to a datum line is the sum of the elevation z, the flow depth y, and the velocity head $v^2/2g$, where v is the mean velocity of flow and g is gravitational acceleration (9.8 m/s^2). The energy is represented by the *energy grade line* or *energy line*, and the loss of energy resulting from water flowing from section 1 to section 2 is h_f (Chow, 1959). The depth of flow, the discharge, and the slopes of the channel bottom and the water surface are interdependent. The flow parameters in Figure 3.1 form the basis for procedures such as step-backwater calculations that are used to estimate discharge from paleostage indicators (Supplemental Section 3.2.1).

Water flowing in an open channel is subject to gravity and friction. External friction results from the channel boundaries. The portion of the flow closest to the boundaries typically moves more slowly than portions of the flow further away from the boundaries as a result of external friction. External friction can vary widely in natural channels because of the presence of an irregular surface over individual grains, the presence of bedforms, bends, changes in channel width, and other factors, as discussed in more detail in Section 3.1.3.

Internal friction results from eddy viscosity (Chow, 1959). *Viscosity* represents the resistance of a fluid to deformation. The *molecular* or *dynamic viscosity*, μ, is the internal friction of a fluid that resists forces tending to cause flow (Robert, 2003). The greater the dynamic viscosity the smaller the deformation within the fluid for a given applied force and the lower the degree of mixing and turbulence. *Kinematic viscosity*, υ, is the ratio of molecular viscosity to fluid density ($\upsilon = \mu/\rho$; typically $\upsilon \sim 1 \times 10^{-6}$ m^2/s). Both types of viscosity decrease significantly as water temperature increases (Fox and McDonald, 1978). The effect of changing density on kinematic viscosity is such that very high suspended sediment concentrations can increase kinematic viscosity (Colby, 1964).

Eddy viscosity is friction within the flow that results from the vertical and horizontal circulation of turbulent eddies. Eddy viscosity expresses the vertical and horizontal transfer of momentum, or exchange between slower and faster moving parcels of water, and varies with position above the bed. The coefficient of eddy viscosity, ε, represents momentum exchange or turbulent mixing (Robert, 2003).

Table 3.1 Types of open-channel flow.

Type of flow	Criterion
Uniform/varied	Velocity is constant with position/ velocity is variable with position
Steady/unsteady	Velocity is constant with time/velocity is varied with time
Laminar/turbulent	$R_e < 500$/ $R_e > 2500$ (transitional flow between)
Subcritical/critical/ supercritical	$F < 1$/$F = 1$/$F > 1$

Considering only the vertical dimension within a channel, eddy viscosity can be expressed as

$$\varepsilon = l^2\, dv/dz \tag{3.1}$$

where l is the *mixing length*—the characteristic distance traveled by a particle of fluid before its momentum is altered by the new environment (Chanson, 1999), dv is change in velocity, and dz is change in height above the bed (Robert, 2003). The mixing length represents the degree of penetration of vortices within the flow and depends on distance from the boundary. (A vortex is the rotating motion of water particles around a common center (Lugt, 1983).) The mixing length is assumed to be

$$l = \kappa z \tag{3.2}$$

where κ is the von Karman constant, which is 0.41 in clear water flowing over a static bed (Robert, 2003). Understanding of eddy viscosity is important because eddies allow dissolved or particulate material carried in the water to spread throughout the flow field.

Open-channel flow is classified into types based on four criteria. The classifications segregate based on properties important to natural processes (Table 3.1). *Uniform flow* occurs when the depth of flow is the same at every section of the channel (Figures 3.1a, 3.2), and is very rare in natural channels. Flow is varied if the depth changes along the length of the channel, and can be either *rapidly varied* (Figure 3.2) if the depth changes abruptly over a

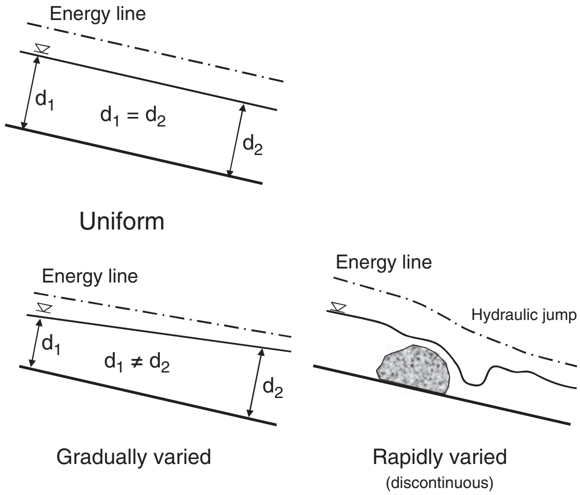

Figure 3.2 Illustration of differences between uniform, gradually, and rapidly varied flow. Longitudinal view of a short channel segment, with flow from left to right. *d* is flow depth. Figure courtesy of David Dust.

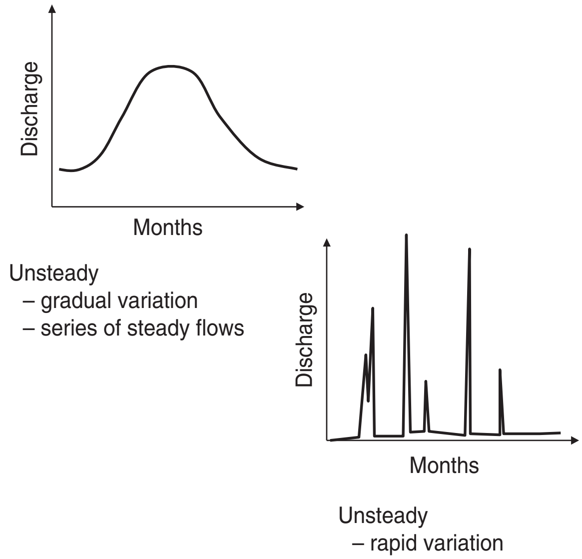

Figure 3.3 Idealized hydrographs for different types of unsteady flow. Figure courtesy of David Dust.

comparatively short distance, or *gradually varied* (Figures 3.1b, 3.2).

Uniform flow can be steady or unsteady. In *steady flow*, the local depth does not change or can at least be assumed to be constant during the time interval under consideration. Like uniform flow, steady flow is rare in natural channels. *Unsteady flow* can occur as gradual variation among a series of steady flows, or as rapid variation (Figure 3.3).

Most natural flows are varied and unsteady, although the variations in time and space can be gradual. Although uniform flow was traditionally assumed for simplicity when calculating hydraulic parameters in natural channels (Chow, 1959), many applications now assume gradually varied flow.

The state of open-channel flow is governed by the effects of viscosity and gravity relative to the inertial forces of the flow. The effect of viscosity relative to inertia can be represented by the *Reynolds number*

$$R_e = \frac{vR\rho}{\mu} \tag{3.3}$$

where v is velocity, R is hydraulic radius (cross-sectional area divided by the wetted perimeter; wetted perimeter is the wetted length of bed and banks at the cross section), ρ is mass density, and μ is dynamic viscosity.

The Reynolds number can be used to differentiate laminar, turbulent, and transitional flow. In *laminar flow*, the viscous forces are sufficiently strong relative to the inertial forces that viscosity exerts a significant influence on flow behavior. Each fluid element moves along a specific path with uniform velocity and no significant mixing between adjacent layers. When the velocity or hydraulic radius exceeds a critical value, the flow becomes *turbulent*. The viscous forces are weak relative to the inertial forces. The fluid elements follow irregular paths and mixing occurs, involving transfer of momentum by large-scale eddies.

The transition from laminar to turbulent depends on the energy dissipation within the fluid: high viscosity equates to more energy necessary to mix the flow. A smooth or glassy water surface in a channel does not mean that the flow is laminar. The flow in most channels is turbulent, but laminar motion can persist in a very thin layer (typically <1 mm thick) next to the channel boundary that is known as the *laminar sublayer* (Figure 3.4).

The Reynolds number provides a useful indicator of the magnitude of mixing within a natural channel.

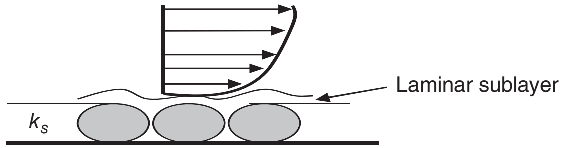

Figure 3.4 Side view of vertical velocity distribution in a channel (flow from left to right). Arrows indicate relative velocity and k_s indicates roughness height. Figure courtesy of David Dust.

The flow is typically turbulent when $R_e > 2300$, and laminar when $R_e < 2300$, although there is no single value of R_e at which the flow changes from laminar to turbulent (Fox and McDonald, 1978).

The effect of gravity on the state of flow is represented by the ratio of inertial forces to gravity forces, as given by the Froude number, F

$$F = \frac{v}{\sqrt{gd}} \tag{3.4}$$

where d is the hydraulic depth, defined as the cross-sectional area of the water normal to the direction of flow in the channel divided by the width of the free surface (Chow, 1959).

When $F = 1$, $v = \sqrt{gd}$, and the flow is in a critical state. When $F < 1$, $v < \sqrt{gd}$, and the flow is *subcritical*. When $F > 1$, $v > \sqrt{gd}$, and the flow is *supercritical*. *Critical flow* can occur when the flow is constricted or subject to a substantial increase in bed slope and passes through a state of minimum specific energy ($v/\sqrt{gd_c} = 1$, where d_c is critical hydraulic depth), as it funnels through or drops over the channel contraction or slope break (Section 3.1.2).

Subcritical flow has lower velocity and greater depth for a given channel configuration and discharge, and is stable and persistent. In a sense, subcritical flow is the usual flow state. Supercritical flow is inherently unstable and changes in channel bed gradient or cross-sectional area can cause the flow to become subcritical, with implications for erosion and deposition along the channel, as discussed in the next section.

3.1.2 Energy, flow state, and hydraulic jumps

As noted earlier, water flowing downhill within a channel is converting potential energy ($PE = mgh$, for water mass m and height h above a given datum) to kinetic energy ($KE = \frac{1}{2} mv^2$), some of which is dissipated at the microscale of turbulence and the molecular scale of heat. A *streamline* is a visual representation of the flow field in the form of a line drawn in the flow field so that, at a given instant of time, the streamline is tangent to the velocity vector at every point in the flow field (Fox and McDonald, 1978). In other words, streamlines indicate the paths of flow, and there can be no flow across a streamline. The total energy in any streamline passing through a channel section can be expressed as the total head H, which is equal to the sum of the elevation above a datum, the product of the depth below the water surface d and the cosine of the bed angle θ, and the velocity head (Figure 3.1)

$$H = z + d\cos\theta + \frac{\alpha v^2}{2g} \tag{3.5}$$

For channels of small slope, $\theta \sim 0$, and the $\cos\theta \sim 1$. Because velocity is not uniformly distributed over a channel section, the velocity head of open-channel flow is typically greater than the value of $v^2/2g$. The true velocity head is expressed as $\alpha v^2/2g$, where α is the energy coefficient and typically varies from about one to two for natural channels (Chow, 1959). The value is higher for small channels and lower for large, deep channels. Every streamline passing through a cross section will have a different velocity head because of the nonuniform velocity distribution in real channels, but the velocity heads for all points in the cross section can be assumed to be equal in gradually varied flow (Chow, 1959).

The line representing the elevation of the total head of flow is the *energy line*, and the slope of the energy line is known as the *energy gradient* S_f (Chow, 1959). The water-surface slope is S_w and the slope of the channel bed is $S_o = \sin\theta$, where θ is the slope angle of the channel bed. In uniform flow,

$S_f = S_w = S_o = \sin\theta$ (Chow, 1959). Based on the principle of the conservation of energy, the total energy head at upstream section 1 (Figure 3.1) should equal the total energy head at downstream section 2 plus the loss of energy h_f between the two sections in parallel or gradually varied flow

$$z_1 + d_1 + \frac{\alpha_1 v_1^2}{2g} = z_2 + d_2 + \frac{\alpha_2 v_2^2}{2g} + h_f \quad (3.6)$$

This is known as the *energy equation*. When $\alpha_1 = \alpha_2 = 1$ and $h_f = 0$, this becomes the Bernoulli equation, which holds in friction-free flow

$$z_1 + d_1 + \frac{v_1^2}{2g} = z_2 + d_2 + \frac{v_2^2}{2g} = \text{constant} \quad (3.7)$$

Specific energy in a channel section is the energy per kilogram of water at any section of a channel measured with respect to the channel bottom (Chow, 1959). Using Equation 3.5, with $z = 0$, specific energy E is

$$E = d\cos\theta + \frac{\alpha v^2}{2g} \quad (3.8)$$

For a channel of small slope and $\alpha = 1$,

$$E = d + \frac{v^2}{2g} \quad (3.9)$$

so that specific energy is equal to the sum of the flow depth and the velocity head. Because $v = Q/A$, $E = d + Q^2/2gA^2$. Therefore, for a given channel section and discharge, E is a function of only the depth of flow (Chow, 1959). A *specific energy curve* is a plot of E and d for a given channel section and discharge (Figure 3.5).

For a given specific energy, two possible depths exist, the low stage and the high stage, each of which is an alternate depth for the other. At point C on the curve, known as *critical depth* d_c, the specific energy is a minimum, which corresponds to the critical state of flow. At greater depths, the velocity is less than the critical velocity for a given discharge and flow is subcritical. At lower depths, the flow is supercritical.

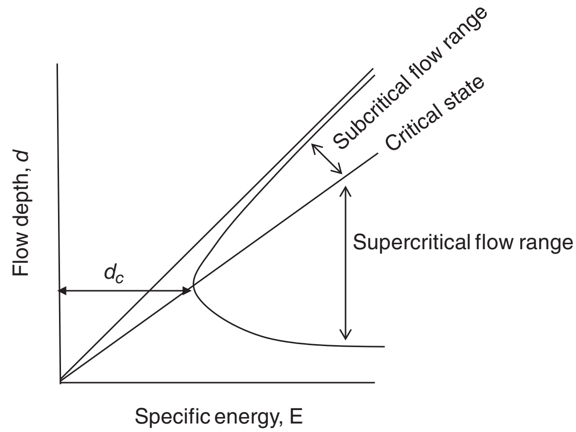

Figure 3.5 Specific energy curve; d_c is critical flow depth.

Specific energy changes as discharge changes, as shown in Figure 3.5. The criterion for critical flow is that the velocity head equals half the hydraulic depth ($v^2/2g = d/2$), assuming that flow is parallel or gradually varied, the channel slope is small, and the energy coefficient equals one (Chow, 1959). Otherwise, $\alpha v^2/2g = (d\cos\theta)/2$.

A flow at or near the critical state is unstable because a minor change in specific energy will cause a major change in depth (Chow, 1959). Consequently, extended areas of critical or supercritical flow are relatively uncommon in natural channels. The flow is likely to return to a subcritical state via a hydraulic jump.

When a rapid change in the flow depth from low stage to high stage occurs, the water surface rises abruptly in a *hydraulic jump* (Figure 3.6). This is common where a steep channel slope abruptly decreases. An *undular jump* forms when the change in depth is small and the water passes from low to high stage through a series of undulations that gradually diminish in size (Chow, 1959). A *direct jump* occurs at a large change in depth and involves a relatively large amount of energy loss through dissipation in the turbulent flow in the jump.

The geomorphic significance of a hydraulic jump is that the jump creates intense turbulence and large kinetic energy losses. Flow velocity decreases substantially across the jump and sediment can be deposited downstream, stabilizing the position of

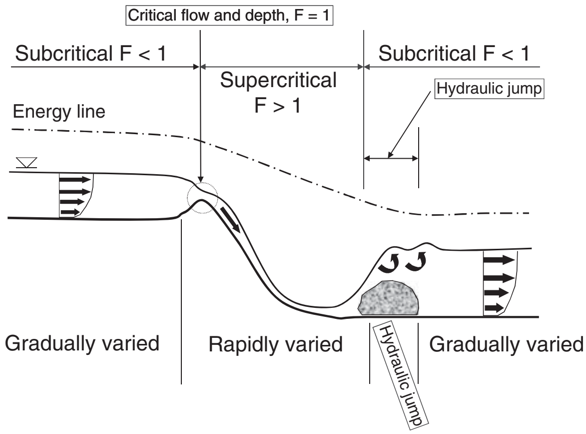

Figure 3.6 Longitudinal view of a hydraulic jump and the associated velocity distributions and flow conditions in channel segments immediately upstream and downstream from the jump. Flow from left to right. F is Froude number. Length of arrows indicates relative velocity. The boulder at the change in slope can localize the jump, but is not necessary for the jump to occur. The abrupt change in slope can be sufficient to create a hydraulic jump. Figure courtesy of David Dust.

the jump (Carling, 1995). If the channel boundaries are adjustable under a given flow, sites of supercritical or critical flow are likely to be preferentially eroded so that the channel cross section enlarges or gradient declines until subcritical flow occurs (Kieffer, 1989; Grant 1997). Natural channels thus exhibit feedbacks between flow energy and channel geometry that tend to maintain subcritical flow (Grant, 1997).

3.1.3 Uniform flow equations and flow resistance

Both velocity distribution within a channel and average velocity are sensitive to boundary roughness that retards flow. Velocity can be directly measured with a velocity meter, and the measured value can be used to calculate flow resistance. Or flow resistance coefficients can be used to estimate velocity under conditions in which direct velocity measurements are not feasible.

Water flowing down an open channel encounters resistance that is counteracted by gravity acting on the water in the direction of motion. Uniform flow occurs when the resistance is balanced by gravity forces. Velocity is related to flow resistance or boundary roughness using one of the three commonly applied uniform flow equations, the Chézy, Darcy–Weisbach, and Manning equations. Because steady uniform flow is rare in natural channels, the results obtained by applying these equations are approximate and general (Chow, 1959). The equations continue to be widely used because they are relatively simple to apply and provide satisfactory approximations for many natural channels.

The French engineer Antoine de Chézy developed the first uniform flow formula

$$v = C\sqrt{RS} \tag{3.10}$$

where v is mean velocity, R is hydraulic radius, S is the slope of the energy line S_f, commonly approximated as water-surface slope S_w or bed slope S_o, and C is a factor of flow resistance that represents the ratio between the driving force (RS) and the velocity sustained in the presence of frictional resistance to flow. Secondary equations developed for calculating the value of C typically rely on an empirically determined coefficient of roughness.

Henry Darcy, Julius Weisbach, and other nineteenth-century engineers developed the approach now known as the Darcy–Weisbach equation

$$v = \left(\frac{8gRS}{f}\right)^{\frac{1}{2}} \tag{3.11}$$

The Darcy–Weisbach equation was developed primarily for flow in pipes, but the form of the equation stated earlier can be applied to uniform and nearly uniform flows in open channels. The variable f is sometimes preferred over other friction factors because it is nondimensional and has a more sound theoretical basis.

Natural channels have irregular boundaries that can be more difficult to characterize in terms

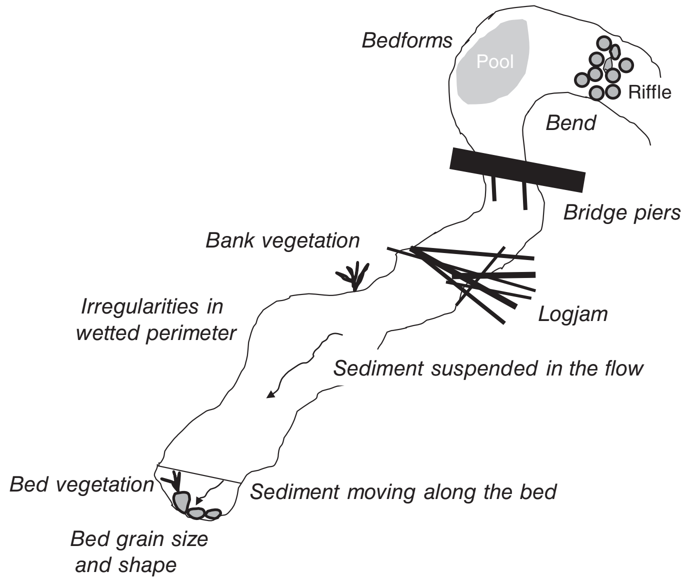

Figure 3.7 Schematic illustration of some of the sources of resistance (named in italicized text) that are lumped together in resistance coefficients such as *C*, *f*, or *n*. Sources of resistance in natural channels include boundary irregularities, vegetation along the channel boundaries, sediment forming the bed and banks, and sediment in transport, obstructions, and channel bends.

of resistance than flow in pipes. The Gauckler–Manning empirical formula was developed in the late nineteenth century by French engineer Philippe Gauckler, and subsequently modified by Irish engineer Robert Manning, to describe the relation among velocity and flow resistance in natural channels. The formula was subsequently modified to its present form

$$v = \frac{R^{2/3} S^{1/2}}{n} \tag{3.12}$$

where n is the coefficient of resistance, typically known as Manning's n. This is the most widely used formula for uniform flow.

The flow resistance factors C and n relate to one another as

$$C = \frac{1}{n}\left(\frac{D_H}{4}\right)^{\frac{1}{6}} \tag{3.13}$$

where D_H is the hydraulic diameter, which is four times the hydraulic radius R (Chanson, 1999).

Determining an n value—or a value for C or f—is the most difficult part of applying these equations, in part because n incorporates so many forms of roughness and flow resistance. Each of the resistance coefficients listed earlier includes multiple sources of resistance (Chow, 1959) (Figure 3.7). Values of n are commonly quite variable in space and in time because n value tends to decrease as stage and discharge increase, although the rate of change in n values can be nonlinear because of interactions between the flow and the channel boundaries. The most common approach is to visually estimate the n value using standard tables (Chow, 1959) or comparisons to field-measured values from diverse sites (Barnes, 1967). Numerous empirical equations also exist for estimating n values for particular types of channels (Supplemental Section 3.1.3).

A key consideration in applying these uniform flow equations is that many natural channels have local accelerations and decelerations associated with protruding grains or bedforms, so that bed, water surface, and energy slopes may not be equal and uniform flow assumptions are not met. Under these circumstances, a uniform flow resistance equation can

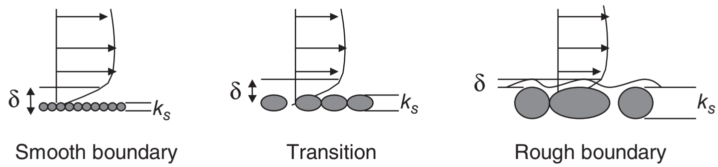

Figure 3.8 Hydraulically smooth and rough boundaries, showing k_s roughness height and laminar sublayer δ. In these beds of uniform grain size, k_s is effectively the grain diameter. In beds with a range of grain sizes, k_s is better approximated by some multiplier of a characteristic grain size (e.g., $3.5D_{84}$), as explained in the text. (From Julien 1998, Figure 6.2.)

be used to relate reach-averaged velocity to reach-averaged depth and bed or water-surface slope if slope is measured over multiple obstacles and bedforms, starting and finishing at sections with the same depth and velocity (Ferguson, 2012). Otherwise, S must be the energy line. The existence of protruding grains and bedforms also makes estimation of R very difficult. The most common approximation is to use a cross-sectionally averaged or reach-averaged value of R in the flow resistance equations.

The Chézy, Darcy–Weisbach, and Manning equations are based on the recognition that the velocity of water flowing in a channel is influenced by the channel boundary roughness. Within the *boundary layer*, velocity varies according to distance from the channel surface. Outside the boundary layer, the velocity distribution is practically uniform (Chow, 1959). In many natural channels, the boundary layer extends to the water surface. As noted in Section 3.1.1, a laminar sublayer can be present at the base of the boundary layer. The top surface of the laminar sublayer corresponds to the transitional zone from laminar to turbulent flow and gives way to the turbulent boundary layer, in which the vertical velocity distribution is approximately logarithmic. The thickness of the laminar sublayer, δ, is defined by

$$\delta = 11.6\upsilon/v^* \tag{3.14}$$

where υ is kinematic viscosity and v^* is shear velocity, which has the dimensions of velocity and is determined from shear stress τ_0

$$v^* = \sqrt{(\tau_0 \rho)} \tag{3.15}$$

The thickness of the laminar sublayer decreases with an increase in shear stress as turbulence penetrates closer to the bed (Richards, 1982; Robert, 2003) but, as noted earlier, the sublayer is typically <1 mm thick.

The thicknesses of the laminar and turbulent portions of the boundary layer partly depend on the characteristics of the boundary roughness. The effective height of the irregularities forming the rough boundary is the *roughness height* k_s (Figure 3.8), and the ratio of the roughness height to the hydraulic radius, k_s/R, is the *relative roughness* (Chow, 1959). If k_s is only a small fraction of the thickness of the laminar sublayer, the surface is hydraulically smooth. If the effects of the roughness elements extend beyond the laminar sublayer, the surface is hydraulically rough and the velocity distribution depends on the form and size of the roughness projections. Because the laminar sublayer is commonly extremely thin, and most natural channels have individual grains and other features that extend beyond this distance into the flow, natural channels have hydraulically rough surfaces.

The turbulent boundary layer is of most interest for river processes because it is within this layer that velocity is measured, shear stress is estimated, and sediment transport is linked to hydraulic parameters (Robert, 2003). The *law of the wall* describes the variation of velocity with height above the bed surface within the turbulent boundary layer, and is used to derive shear stress and roughness height from measured velocity profiles (Robert, 2003)

$$v_z = 2.5v^* \ln(z/z_0) \tag{3.16}$$

where v_z is the mean one-dimensional (1D) velocity at a height z above the bed, z_0 is the projected height above the bed at which velocity is zero, and the constant 2.5 is equal to $1/\kappa$.

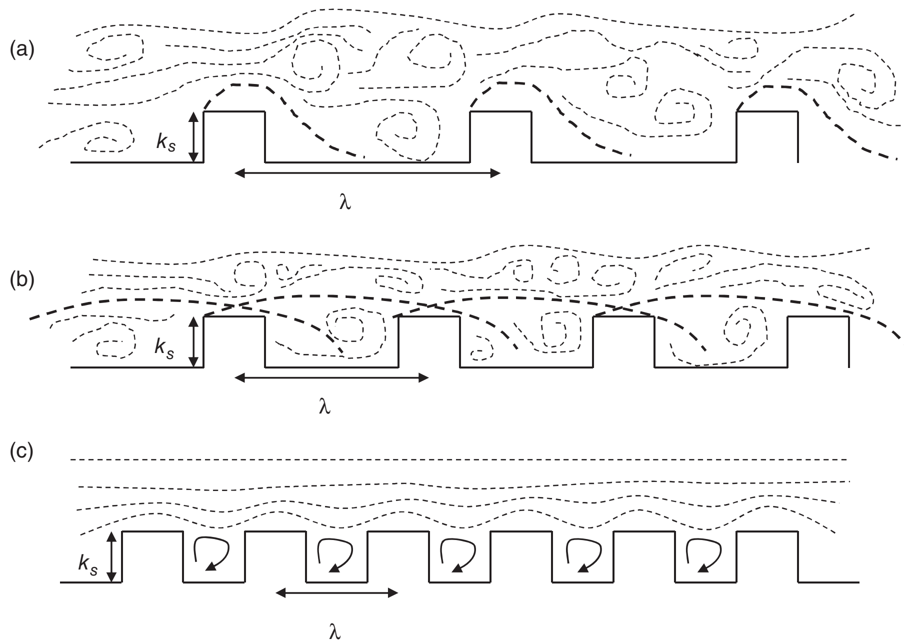

Figure 3.9 Longitudinal view of differing types of bed roughness configurations. (a) Widely spaced roughness elements create isolated-roughness flow in which the wake and vortex at each element develop and dissipate before the next element is reached. (b) More closely spaced roughness creates wake-interference flow in which the wake and vortex from each element interfere with those developed at the next roughness element downstream. This creates intense turbulence and complex vorticity. (c) Roughness elements sufficiently close together to create flow that skims the crest of each element is quasi-smooth flow. Each depression between elements has extremely low velocity and a stable eddy. (From Chow, 1959, Figure 8-4, p. 197.)

The law of the wall can be integrated from $z = z_0$ to $z = d$, where d is flow depth, to calculate mean 1D velocity at a vertical section via the Keulegan equation

$$v/v^* = 2.5\ \ln(d/k_s) + 6 \tag{3.17}$$

where v is the average velocity at a vertical, v^* is shear velocity, and k_s is roughness height (Robert, 2003). The primary challenge is the determination of k_s. With densely packed grains of uniform size and shape and a flat, immobile bed, k_s is approximately the median diameter of the bed sediment (Robert, 2003). With poorly sorted, heterogeneous bed sediment and bedforms, k_s can vary widely. Commonly used, empirically based approximations include $3.5D_{84}$, $6.8D_{50}$, and $2D_{90}$, where D_x is the grain size for which x percentage of the cumulative grain size distribution is finer. Average values of the roughness dimensions can be used to represent surfaces of variable roughness, as in channels with a wide range of grain sizes, although much effort has been devoted to determining how to most accurately characterize an average.

The loss of energy in turbulent flow over a rough boundary results from the formation of wakes behind each roughness element, so the longitudinal spacing λ of the roughness elements is particularly important. *Isolated-roughness flow* occurs where individual roughness elements are sufficiently far apart that the wake and vortex at each element are completely developed and dissipated before the flow reaches the next element (Figure 3.9). Roughness results from form drag on the roughness elements and friction drag on the wall surface between elements. *Wake-interference flow* occurs where the roughness elements are sufficiently closely spaced for the wake and vortex at each element to interfere with the wake and vortex developed at the next element. This results in intense and complete vorticity and turbulence mixing, as well as the highest values of flow resistance. The transition from isolated-roughness to wake-interference flow occurs when the ratio of downstream spacing between obstacles (L) to height of the obstacles (H) is approximately 9–10 (Wohl and Ikeda, 1998). *Skimming flow* occurs where the roughness elements are so closely spaced that the flow skims the crests of the elements. Large roughness projections are absent and the surface acts hydraulically smooth (Chow, 1959). Isolated-roughness elements can inhibit entrainment of surrounding grains and affect bedload motion, and these effects become more pronounced as roughness elements are more closely spaced on the bed.

As noted earlier, resistance to flow along the bed in most natural channels results from diverse features. These features are commonly divided into three categories (Griffiths, 1987; Robert, 2003). The friction created by individual grains within heterogeneous bed sediments is designated *grain resistance* (or skin or grain friction or grain roughness). Friction from individual grains organized into bedforms such as dunes, pools and riffles, or steps and pools is known as *form resistance* (or form roughness or form drag). Friction also results from pressure and viscous drag on sediment in transport above the bed surface (Griffiths, 1987). Sometimes a distinction is also drawn between relatively small bedforms and large bed undulations that create *spill resistance* because of a vertical drop in the water surface (Leopold et al., 1960).

An important consideration is that roughness is related to uniform flow equations. Anything that causes the water surface to shift to gradually or, especially, rapidly varied flow (such as a large bedform or a logjam) is an obstruction that does not formally fall under the category of form roughness as used by hydraulic engineers. In practice, however, any large obstacle to flow is lumped under the category of form roughness in most geomorphic discussions of flow resistance in natural channels, and roughness is used generically to refer to various forms of flow resistance. In the discussion that follows here, "resistance" is used to include roughness elements and elements that cause the water surface to shift to varied flow. (In engineering hydraulics, roughness refers to a coefficient that represents the effects of boundary shear stress in uniform flow equations, resistance is a force with a vector (directional roughness), and energy dissipation refers to everything that retards flow, including flow transitions such as hydraulic jumps, eddies, and grain and form resistance.)

Einstein and Banks (1950) used flume experiments to demonstrate that grain, f_g, and form, f_b, resistance could be additive

$$f = f_g + f_b \tag{3.18}$$

Grain resistance is negligible in channels formed in sand-sized and finer sediment. Individual particles do not protrude into the flow above the rest of the bed and, except at very shallow flow depths, individual particles influence only a small portion of the flow. Grain resistance can become quite substantial in channels with larger grains, particularly where the grains are poorly sorted and are large relative to flow depth. Poorly sorted substrates with a wide range of grain sizes allow large grains to protrude into the flow above the rest of the bed. Individual grains may create the greatest resistance in shallow, steep, bouldery headwaters, whereas bedforms such as step–pool and pool–riffle sequences (Section 4.3.2) formed in gravel dominate flow resistance in the middle segments of drainage networks (Prestegaard, 1983). Other sources of flow resistance become more important in the deeper, low-gradient segments of drainages.

The importance of grain resistance can be expressed using R/k_s when k_s is a function of grain size. This ratio is known as the *relative grain submergence*, commonly expressed as R/D_{84}, and is the inverse of relative roughness. Relative grain submergence can be used to distinguish large-scale roughness ($0 < R/D_{84} < 1$) in which individual grains protrude a substantial distance into the flow, intermediate-scale roughness ($1 < R/D_{84} < 4$) and small-scale roughness ($R/D_{84} > 4$) in which grains protrude above the bed only a small proportion of flow depth (Bathurst, 1985) (Figure 3.10). Relative form submergence, R/H, where H is bedform

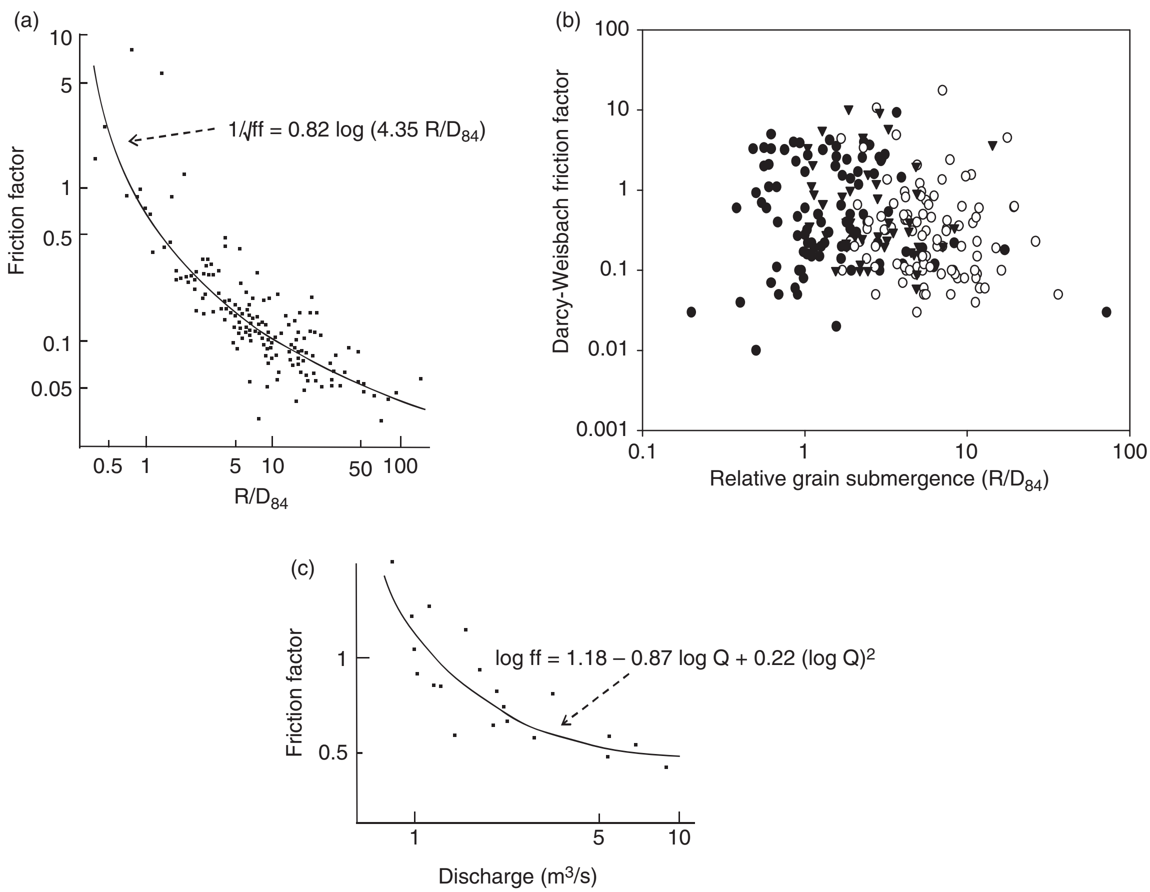

Figure 3.10 The relationship between Darcy–Weisbach friction factor and relative grain submergence (R/D_{84}) differs between (a) low-gradient alluvial channels with sand bedforms in which friction factor decreases as grain roughness is drowned out (from Knighton, 1998, Figure 4.2A) and (b) high-gradient alluvial channels with cobble- to boulder-beds in which the range of values for friction factor is relatively constant (from Wohl and Merritt, 2008, Figure 9). Step–pool channels are solid circles, pool–riffle channels are open circles, and plane-bed channels are solid triangles. (c) Friction factor tends to decrease with discharge, as shown in this plot of data from the River Bollin, UK (from Knighton, 1998, Figure 2B), although the rate of decrease varies among channels as a function of the sources of roughness.

amplitude, can be similarly used to express the importance of form roughness (Wohl and Merritt, 2008).

The phrases "grain resistance" and "form resistance" can be misleading in that individual large grains actually create form resistance. Grain resistance is the cumulative effect of many individual grains that retard water flowing over the bed in a uniform manner, without accelerations or decelerations and with parallel streamlines and the same mean depth and velocity (Ferguson, 2013). Resistance results from viscous shear over the grain surfaces and local pressure gradients around the grains. Large grains that protrude into the flow create form resistance as larger-scale flow acceleration and deceleration generate large turbulent eddies in which energy is dissipated by viscous forces (Ferguson, 2013). Form resistance thus includes the effects of drag on large roughness elements such as protruding boulders or logs, the spill losses over steps or riffles, standing waves around protruding obstacles, and transverse accelerations caused by sharp changes in flow direction (Ferguson, 2013). Grain and form resistance can be very difficult to distinguish in many natural channels.

A widely used approach (Robert, 2003) for partitioning resistance in natural channels has been to select an appropriate value of k_s for grain resistance and use this value to calculate f_g via the equation

$$1/\sqrt{f_g} = 2.11 + 2.03\log_{10}(d/k_s) \qquad (3.19)$$

Grain resistance is then subtracted from total resistance, f, to estimate form resistance. However, flume studies that combined grain resistance and form resistance with resistance from individual instream wood pieces and jams indicate the existence of substantial interaction effects between resistance components. Consequently, a simple additive approach can be inaccurate (Wilcox et al., 2006).

An alternative to assuming that any portion of f not included in f_g is f_b involves directly calculating f_b

$$f_b = 0.5C_d h\rho v^2 \qquad (3.20)$$

where C_d is a drag coefficient that varies with obstacle shape and h is the height of bed undulations associated with the bedforms (Robert, 2003). Quantifying C_d is difficult, however, because form drag is influenced by the pressure field, and thus the velocity field, above the bedforms. The *pressure field* is the instantaneous water pressure exerted on a given surface, such as the bed, within a control volume of fluid, expressed as a function of three spatial coordinates and time. The *velocity field* is the instantaneous velocity of the center of gravity of a volume of fluid instantaneously surrounding a point in the flow (Fox and McDonald, 1978). In other words, C_d for a given obstacle is not constant, but varies with discharge and with the configuration of the channel geometry and bed resistance in the immediate vicinity of the obstacle, as these influence local velocity and pressure distributions. Quantifying and partitioning flow resistance, particularly in coarse-grained channels, continues to be an elusive goal.

Although more attention has been devoted to bed resistance, bank resistance can also exert an important influence on total resistance and the distribution of hydraulic forces (Wohl et al., 1999; Kean and Smith, 2006a, 2006b). Bank resistance results from: individual grains; bank vegetation; irregularities such as those produced by slumping of bank material or the presence of vegetation; and repetitive variations in the channel planform such as bends. Examining Lost Creek, a small (4 m wide), gravel-bed channel with rough banks (±20 cm amplitude), Kean and Smith (2006a) found that additional flow resistance created by drag on bank topographic features substantially reduced near-bank velocity and shear stress. Neglecting these effects in Lost Creek resulted in a 56% overestimate of discharge. Enhanced bank resistance caused by vegetation can alter the spatial distribution of velocity, shear stress, and sediment deposition, and thus channel geometry (Griffin et al., 2005; Gorrick and Rodríguez, 2012). Friction associated with woody riparian vegetation on the lateral boundaries of the sand-bed Rio Puerco in New Mexico, USA, (average channel width ~11 m) reduced perimeter-averaged boundary shear stress by almost 40% and boundary shear stress in the channel center by 20% (Griffin et al., 2005). The effects of bank resistance may be more important along channels formed in finer sediment, such as sand, than in more coarse-grained cobble- to boulder-bed channels in which bed resistance is very large (Yochum et al., 2012).

Momentum equations define the hydrodynamic forces, such as drag, exerted by flow. Houjou et al. (1990) proposed that the velocity field in flows for which lateral shear is an important factor can be calculated using a momentum equation in the form of

$$-\rho g S = \frac{\partial}{\partial z}\left(\kappa\frac{\partial u}{\partial z}\right) + \frac{\partial}{\partial x}\left(\kappa\frac{\partial u}{\partial x}\right) \qquad (3.21)$$

where ρ is fluid density, g is acceleration due to gravity, S is the downstream water-surface gradient, κ is kinematic eddy viscosity, z is the coordinate normal to the bed, x is the transverse coordinate, and u is the downstream velocity component. Houjou et al. (1990) explained that, in a channel with a rectangular cross section, flow structure reflects width–depth ratio and the ratio of wall and bed resistance.

Equation 3.21 permits computation of velocity and stress fields affected by both the channel bed and banks.

Water flowing through a bend experiences a transverse force proportional to v^2/r, where r is the radius of curvature. This adds to the total flow resistance in sinuous channels (Ferguson, 2013). Total head loss along a reach scales with R/r^2 (Chang, 1978). Radius of curvature increases more rapidly than v or R as river size increases, so bend losses decrease in importance with river size and are insignificant in large rivers for which banks are readily erodible and in which regular meanders can develop (Ferguson, 2013).

Supplemental Section 3.1.3 provides more discussion of resistance estimates and partitioning resistance.

3.1.4 Velocity and turbulence

Velocity is a vector quantity, with magnitude and direction, and one of the most sensitive and variable flow properties. Because of the presence of a free surface and friction along the channel boundaries, velocity is not uniformly distributed within a channel. Velocity varies with distance from the bed, across the stream, downstream, and with time (Figure 3.11). At a cross section, the maximum

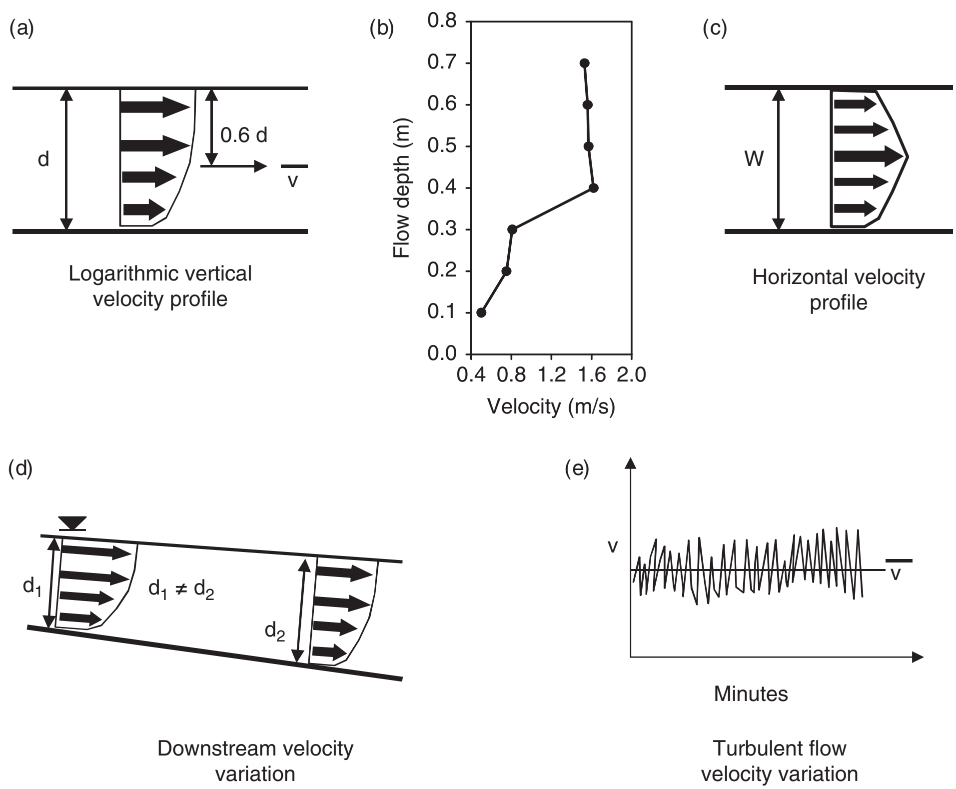

Figure 3.11 Different characteristics of velocity through time and space. (a) Idealized logarithmic vertical velocity profile (flow from left to right, length of arrows indicates relative velocity); d is flow depth, v is mean velocity. (b) Vertical velocity profile from a boulder-bed channel. Abrupt increase in velocity between 0.3 and 0.4 m flow depth reflects top of boulders and transition from highly turbulent flow with low average velocity to rapid flow in upper profile. (c) Horizontal velocity profile looking down on a channel of width w. (d) Longitudinal view illustrating downstream variations in velocity, as reflected in differences within successive vertical velocity profiles. (e) Temporal variations in velocity at short time intervals associated with turbulence. Parts (a), (c), (d), and (e) courtesy of David Dust.

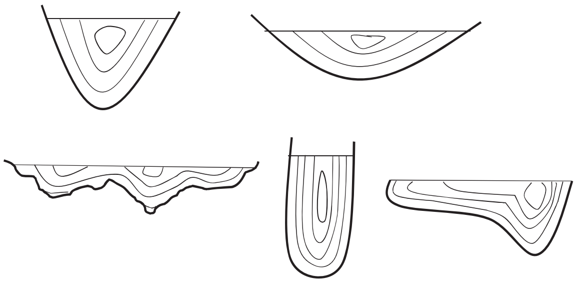

Figure 3.12 Isovelocity lines in differently shaped channels. In each of these channels, the view is looking downstream within the channel. These are idealized distributions: actual distributions are likely to be less symmetrical with respect to cross-sectional boundaries because of upstream effects.

velocity typically occurs just below the free surface. The location of the maximum velocity with respect to distance between the two channel banks depends on the symmetry or asymmetry of the channel (Figure 3.12). Velocity tends to remain constant or increase slightly downstream as resistance from the channel boundaries affects a smaller proportion of the total flow volume. Velocity varies at short time intervals of a few seconds as a result of flow turbulence, and at longer time intervals as a result of roughness of the channel boundaries and fluctuations in discharge during unsteady flow.

As noted earlier, velocity can be characterized in terms of the velocity field. At a given instant in time, the velocity field is a function of the space coordinates x or w (cross stream), y or u (downstream), and z or v (vertical with respect to the bed), so that a complete representation of the velocity field is given by $V = V(x, y, z)$ (Fox and McDonald, 1978). The usual convention is for velocity terms u, v, w to correspond to spatial coordinates x, y, z: hence the use of z for depth or vertical position in some of the equations in this chapter. A flow field described in this manner is *three dimensional* because the velocity at any point in the flow field depends on the three coordinates required to locate the point in space. A flow can be classified as one, two, or three dimensional, depending on the number of space coordinates required to specify the velocity field.

1D flow occurs where velocity is represented by only one dimension, such as velocity along the channel centerline of a long, straight channel with constant cross section, where velocity can be primarily influenced by distance above the channel bed. 1D representations have cross-sectionally averaged velocity along the hydraulic axis that follows the thalweg. From the continuity equation, mean 1D velocity is discharge divided by cross-sectional area ($v = Q/A$).

2D flow occurs when the velocity field is predominantly influenced by two of the space coordinates, such as distance above the bed and from the bank. 2D representations of flow have vertically integrated velocity with only the cross- and downstream components.

The complexity of analysis increases substantially with the number of dimensions of the flow field. Consequently, the simplest case of 1D flow is widely used to provide approximate solutions for evaluating features such as thalweg velocity profile (Fox and McDonald, 1978).

Flow along channels with low gradients and relatively uniform bed material is commonly approximated by a semilogarithmic velocity profile in which velocity varies with distance from the bed (Leopold et al., 1964): this is an example of assuming 1D flow. The profile includes a laminar sublayer, a turbulent boundary layer with logarithmic profile, and an outer layer that deviates slightly from logarithmic (Ferguson, 2007). For channels with rough boundaries or irregular bed material, velocity is proportional to roughness-scaled distance from the bed, rather than just distance from the bed. In effect, the velocity profile becomes segmented, with low velocity flow between larger boulders or below the average surface elevation of the bed, and an abrupt transition to higher velocity flow above the level of the boulders (Figure 3.11b). This two-part velocity profile has been described as s-shaped, although the profile varies substantially in relation to channel boundary roughness. The law of the wall is used to describe the variation of velocity with height above the bed surface in the bottom 20% of fully turbulent flows (Robert, 2003).

Table 3.2 Types of current meters (Clifford and French, 1993b).

Type of meter	Characteristics
Mechanical impellor	Inexpensive; durable; poor frequency response (<1 Hz), 1D data; requires maintenance and recalibration
Electromagnetic	Intermediate cost; robust; good frequency response (5–20 Hz); tolerate particle and other contamination in the flow; provide 1D, 2D, or 3D measurements; frequency response affected by head design
Ultrasonic	Expensive; fragile; sensitive to particle and air bubble contamination; provide 2D and 3D data; excellent frequency response (up to 30 Hz)
Laser velocimeters	Highest cost; sensitive to suspended sediment; produce very high frequency response; 3D data; do not perturb flow being measured; do not function adequately near the streambed

Supplemental Section 3.1.4 provides more information about velocity profiles.

Velocity in natural channels can be measured using current meters or dilution tracers. Point measurements use various types of current meters, including mechanical impellers, electromagnetic and ultrasound meters, and laser velocimeters. Table 3.2 summarizes the characteristics of each type of current meter. One of the limitations of point measurements is the number and spatial density of measurements needed to characterize mean velocity.

Acoustic Doppler current profilers can obtain spatially dense point measurements in a very short period of time, but are very expensive and can be of limited usefulness in very shallow or highly aerated flow. Doppler current meters use the Doppler effect of sound waves scattered back from particles within the water column. The meter generates and receives sound signals. The traveling time of sound waves can be used to estimate distance, and the change in sound wavelength can be converted to velocity.

Dilution tracer techniques using a fluorescent dye or a chemical such as NaCl can also be used to characterize the average velocity for a channel segment (Wohl, 2010). This approach is inexpensive and fast, but provides no information on the distribution of velocity within the measured channel segment.

Supplemental Section 3.1.4 provides more information about velocity measurement techniques.

As water particles move in irregular paths, momentum is exchanged between different portions of the water (Chanson, 1999). *Turbulence* occurs when water parcels flow past a solid surface or past an adjacent water parcel with a different velocity (Clifford and French, 1993a, 1993b). Turbulence appears as irregular velocity fluctuations. Time series of velocity in natural channels commonly reveal intriguing variations in the magnitude of even 1D measurements (Figure S3.1), let alone three-dimensional (3D) flow, and a great deal of human energy is devoted to characterizing the patterns of velocity fluctuations in order to reveal the underlying processes and thus improve understanding of the relations among resistance, hydraulics, and channel adjustment. Understanding of velocity and turbulence is important because of the feedbacks among turbulence, channel morphology, and sediment transport (Figure 3.13) (Leeder, 1983; Buffin-Bélanger et al., 2013).

Channel morphology creates resistance and contributes to the variability in the intensity of turbulent exchanges with the water column. The structure of turbulent flows consists of the mean structure, as reflected in the 1D vertical velocity profile, and the temporal fluctuations of the structure associated with turbulence. Both of these elements influence sediment mobility. Mobile sediments can alter the flow structure and contribute to bedforms, which then in turn influence flow resistance and turbulence.

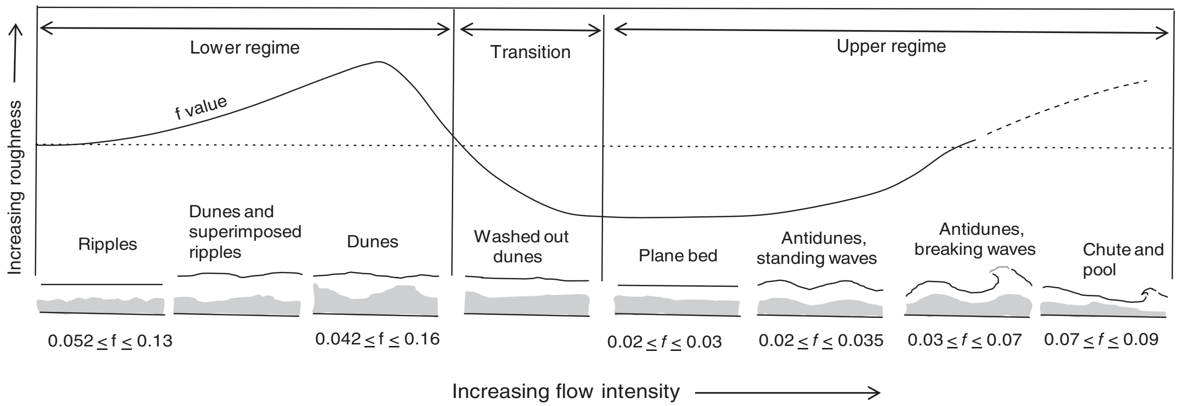

Figure 3.13 Sequence of bedforms that develop with increasing discharge or velocity in mobile-bed channels, and associated flow resistance. Small drawings at the base of the figure are longitudinal views of a segment of bed, indicating the bed profile (gray shading) and the water surface configurations (upper black line). Note that resistance, as indicated by the Darcy–Weisbach friction factor *f*, generally increases going from ripples to dunes in the lower regime, declines dramatically when the dunes wash out to a plane bed, and then increases less in the progression toward chutes and pools within the upper regime, as indicated by the undulating line partway up the y-axis in this figure. (From Simons and Richardson, 1966.)

Turbulence is quantified based on measurements of velocity fluctuations at a point in the flow. The *turbulence intensity* of any of the three velocity components (*x, y, z*) is the average magnitude of the deviation from the mean for a given velocity series (Robert, 2003). Variability around the mean for a normal or near-normal distribution is represented by the standard deviation of the velocity distribution, known as the root-mean-square value, *RMS*, which is used to represent turbulence intensity (Robert et al., 1996; Robert, 2003).

$$RMS_y = \sqrt{\left(\frac{\left(\sum v_y^2\right)}{N}\right)} \qquad (3.22)$$

where v_y is the downstream component of velocity and N is the total number of observations in a given series. The *RMS* value can be computed for any of the three dimensions of velocity and the three expressions can be combined into an index of total turbulence intensity (Robert, 2003).

Flow structures in the vicinity of large obstacles increase the local turbulence intensity in the near-bed region. The greater the value of RMS the greater the turbulence and the greater the potential for sediment entrainment and transport. The turbulent kinetic energy of the flow, *TKE*, is then

$$TKE = 0.5\rho(RMS_x^2 + RMS_y^2 + RMS_z^2) \qquad (3.23)$$

where ρ is water density (Clifford and French, 1993a; Robert, 2003). *TKE* is the mean kinetic energy per unit mass associated with eddies in turbulent flow, and can be conceptualized as the energy extracted from the mean flow by turbulent eddies (Bradshaw, 1985; Robert, 2003). The downstream component of *TKE* is typically dominant in natural channels (Robert, 2003).

Turbulence can also be studied by following the trajectories of tracers traveling with the flow. This *flow visualization* approach uses hydrogen bubbles, colored fluids, or fine particles of neutral submerged density (Buffin-Bélanger et al., 2013). Flow visualization does not provide the quantitative information of Equation 3.22 or 3.23, but does provide insight into the location, size, and structure of turbulence.

Reynolds stresses quantify the degree of momentum exchange at a point in the flow (Robert, 2003). Six Reynolds stresses can be calculated as the product of the negative value of water density ($-\rho$) and the product of the average fluctuations of either a single velocity component (e.g., v_x^2) or two different velocity components (e.g., $v_x v_y$) (Buffin-Bélanger et al., 2013). Like *TKE*, Reynolds stresses are important descriptors of the intensity of turbulent exchanges and are linked to sediment transport (Buffin-Bélanger et al., 2013). Values of RMS, TKE, and Reynolds stresses are each widely used to quantify the turbulence characteristics of flow in natural channels.

Quadrant analysis is commonly used to understand and describe the structure of turbulence in two dimensions. The instantaneous horizontal and vertical velocities are divided into four quadrants based on their deviation from the mean (Robert, 2003). Quadrant I represents positive deviation in the horizontal and vertical. Quadrant II represents negative deviation in the horizontal and positive deviation in the vertical. Quadrant III is negative deviation in both quadrants. Quadrant IV is positive deviation in the horizontal and negative deviation in the vertical. Quadrant II events have slower than average downstream flow velocity and positive vertical flows away from the channel boundary and are known as ejection events or *bursts* because water is ejected from the bed upward into the outer flow (Robert, 2003). Quadrant IV events have greater than average downstream flow velocity and negative vertical flows toward the bed and are known as *sweeps*. Alternate zones of low- and high-velocity water near the bed create *low-speed streaks* that are relatively narrow zones of low-velocity water near the bed (Robert, 2003). The spacing of low-speed streaks reflects the shear velocity and fluid kinematic viscosity. Low-speed streaks culminate in ejections upward from the bed (bursts). To preserve continuity, bursts are followed by high-velocity outer layer flow penetrating the near-bed flow (sweeps). Sweeps lose momentum as they impact the bed and diffuse laterally (Best, 1993; Robert, 2003). This sequence of bursts and sweeps exerts an important control on sand-bed bedforms such as ripples and dunes by moving sand grains and influencing the size of bedforms (Best, 1992, 2005) (Section 4.3.1).

Bursts and sweeps occur only a small fraction of the time, but cause most of the momentum exchange and can strongly influence sediment transport (Robert, 2003). This is the key point regarding turbulence in natural channels, and one of the reasons so much effort is focused on quantifying turbulence. Numerous studies of sediment mobility and associated channel stability indicate that turbulent fluctuations exert a stronger influence on sediment dynamics, particularly initiation of motion, than does mean velocity, especially in channels with beds composed of sediment coarser than sand size.

Turbulent boundary layers include various types of vortices and irregular, but spatially and temporally repetitive, flow patterns known as *coherent flow structures*. Coherent flow structures are self-perpetuating (Smith, 1996; Robert, 2003). Low-speed streaks alternate with zones of high-velocity water and give rise to coherent flow structures such as horseshoe vortices. *Horseshoe vortices*, which are named for their shape (Figure 3.14), are advected upward from the bed.

A second major group of vortices and coherent flow structures are present over coarse-grained surfaces in which vortices are induced by individual large particles or bedforms (Roy et al., 1999; Robert, 2003) (Figure 3.15). Eddies are shed from the downstream end of obstacles and from the shear layer along the flow separation zone downstream from obstacles. Standing vortices form upstream of individual obstacles (Robert, 2003). Horseshoe vortices form from the flow separation zone downstream from isolated obstacles.

The frequency of eddy shedding downstream from isolated obstacles is expressed by the dimensionless *Strouhal number*, *Str*

$$Str = \frac{(feD)}{v} \tag{3.24}$$

where *fe* is the frequency at which eddies are shed from obstacles, *D* is obstacle size, and *v* is mean flow velocity (Robert, 2003). Large Strouhal numbers (~1) indicate that viscosity dominates fluid

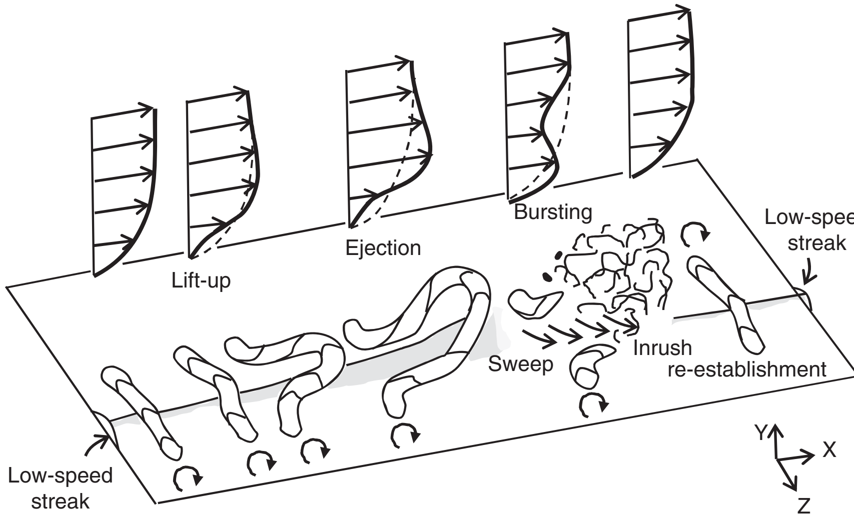

Figure 3.14 Schematic representation of turbulent bursting, according to Allen (1985). Flow from left to right, with length of arrows representing relative velocity. A developing horseshoe vortex lifts, stretches, and bursts. Associated velocity profiles shown at rear (from Bridge, 2003, Figure 2.12a, p. 29).

flow. Low Strouhal numbers ($\leq 10^{-4}$) indicate that the high speed, quasi steady-state portion of the fluid flow dominates. Intermediate Strouhal numbers reflect the formation and rapid shedding of vortices (Sobey, 1982).

Eddy shedding may be the dominant mechanism of energy dissipation in coarse-grained channels (Robert, 2003), and may cause the greater-than-expected flow resistance of coarse-grained channels (Clifford et al., 1992). Pseudo-periodic oscillations in time series of velocity fluctuations may reflect the periodicity at which eddies are shed (Robert, 2003).

Very large roughness elements or an abrupt change in the channel boundary or orientation create *separated flows*, or portions of the channel in which there is no downstream flow (Robert, 2003). Examples of sites inducing flow separation include channel bends, channel expansions, pools, bedforms, and very large grains. The boundary between the lower velocity separation zone and the higher velocity main flow is a zone of rapid change in velocity, intense mixing, and high turbulence intensity known as a *shear layer* (Figure 3.16). The acceleration of flow around the obstacle can allow the turbulent boundary layer to detach or separate from the bed or bank. The *reattachment point* occurs where the turbulent boundary layer reattaches to the bed or bank downstream from the obstacle, and the area between the separation and reattachment points is the zone of *recirculating flow* (Buffin-Bélanger et al., 2013). Zones of recirculating flow are typically sites where finer sediment is deposited as a result of lower velocity. Recirculating flow can also form important aquatic habitat by concentrating organic matter in transport and providing a low-velocity resting area for organisms such as fish. Strong gradients in water properties such as velocity or density where two rivers join can also promote lateral momentum exchanges and large and intense shear layers between the flows (Buffin-Bélanger et al., 2013). The energy dissipated at these boundaries can facilitate sediment deposition.

3.1.5 Measures of energy exerted against the channel boundaries

The force driving a flow, F_d, along a unit length of channel can be expressed as

$$F_d = W \sin\theta \qquad (3.25)$$

where W is the weight of the water and θ is the channel bed slope. Weight equals mass times

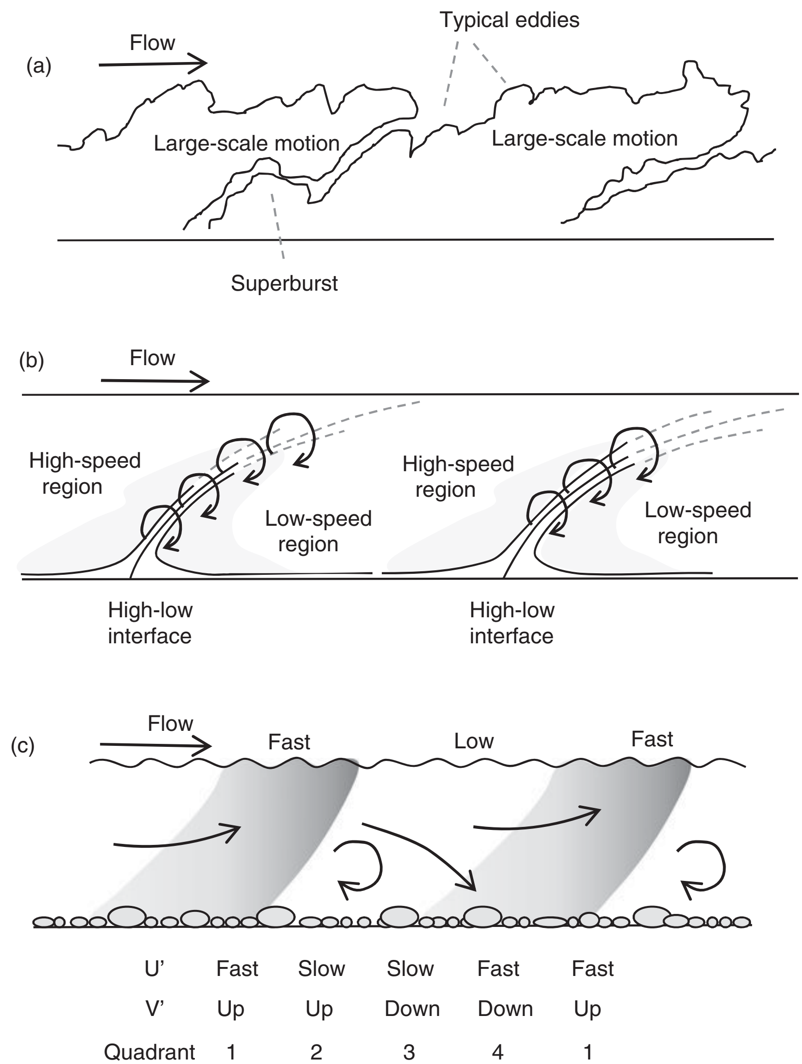

Figure 3.15 Longitudinal views of turbulence. (a) Large-scale flow motions with the presence of typical eddies at their boundaries. (b) High- and low-speed regions within the flow. (c) A sequence of turbulent flows associated with the passage of high- and low-speed wedges. (From Buffin-Belanger et al., 2000, Figure 1.)

gravitational acceleration, g, and $W \sin\theta$ represents the downslope components of mass acted on by g. Mass equals mass density, ρ, and the volume of water, V; (mass $= \rho V$), so

$$F_d = \rho g V \sin\theta \tag{3.26}$$

Stress is force per unit area (A), so the average shear stress exerted by the flow on the bed, τ_0, under conditions of steady uniform flow is

$$\tau_0 = \rho g R \sin\theta \tag{3.27}$$

where R is hydraulic radius. For small slopes, $\sin\theta$ is approximately equal to $\tan\theta$, which is approximated by slope, S, so that

$$\tau_0 = \rho g R S \tag{3.28}$$

This is the basic equation for average *bed shear stress* in natural channels (Robert, 2003). Although steep channels violate the assumption that bed slope approximates $\sin\theta$, Equation 3.28 is commonly applied to steep channels.

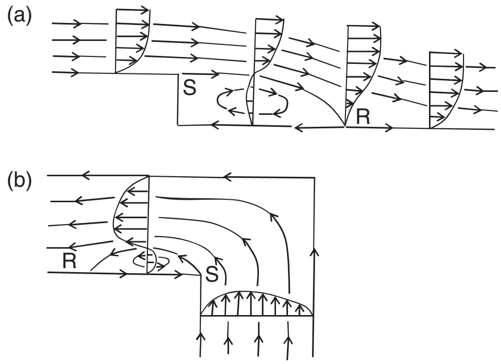

Figure 3.16 Examples of flow separation (S is separation point, R is reattachment point) as indicated by time-averaged streamline patterns at (a) a downward transverse step, as seen in side view, and (b) a sharp bend in an open channel, as seen in planform (from Allen, 1994, Figure 2.24).

Stream power is also commonly used to express energy exerted against the channel boundaries. Stream power is the rate of doing work in transporting water and sediment. Stream power can be expressed as *total power*, Ω

$$\Omega = \gamma QS \qquad (3.29)$$

where γ is the specific weight of water (assumed to be 9800 N/m^3), Q is discharge, and S is channel gradient.

Stream power can also be expressed as *stream power per unit area*, ω

$$\omega = \tau_0 v \qquad (3.30)$$

where τ_0 is bed shear stress and v is velocity.

Finally, stream power can be expressed as *specific stream power*, ω

$$\omega = \Omega/\omega \qquad (3.31)$$

where w is channel width.

Each measure of stream power quantifies the energy available to perform geomorphic work against the channel boundaries in the form of entrainment and erosion.

Spatial variations in stream power are addressed in a model developed by Knighton (1999). Depending on the downstream rates of change for Q and S, the model predicts that total power peaks at an intermediate distance between the drainage divide and the river outlet. Specific stream power is more sensitive to the rate of change in S and peaks closer to the headwaters. The model accurately predicts observed conditions along lowland alluvial rivers (Knighton, 1999), but limited tests for mountain rivers reveal substantial deviations from expected patterns because of local factors that influence Q, S, and w (Fonstad, 2003).

All of the measures of stream power noted earlier are essentially instantaneous values at a point or cross section, and do not account for variations in flow, and thus stream power, through time. Temporal variations in stream power relate mainly to variations in discharge. In one study along the Tapi River, India, the annual peak flood contributed up to 34% of the total energy expended during a monsoon season (Kale and Hire, 2007). Integrating variations in stream power through time and along a channel is necessary to effectively understand the influence of energy expenditure on channel form and process.

3.2 Hydrology

3.2.1 Measuring, indirectly estimating, and modeling discharge

The basic data on river discharge come from gaging stations, typically operated by either government agencies or private companies that require knowledge of water supply. In the United States, for example, the majority of gages are operated by the U.S. Geological Survey, although state agencies also operate many gages. The Environment Agency operates most stream gages in England and Wales. In these countries, discharge data are publicly accessible online at no charge. Government agencies in other countries, such as India, consider stream records sensitive information and will not readily release the data.

Although the continuity equation ($Q = w\ d\ v$) suggests a simple means to measure discharge at a gage, w, d, and v are not measured continuously. Instead, they are initially measured over a range of flows and used to establish a *rating curve* for the cross section. This curve relates Q to *stage* (water-surface elevation), which is in turn related to d, the most readily measured parameter. A rating curve is typically recalibrated periodically to ensure that gradual, progressive changes in cross-sectional geometry do not affect the accuracy of discharge calculations. This method of calculating discharge is most accurate where cross-sectional geometry is relatively simple, without split flow or extensive backwaters, and stable through time. Cross sections with bedrock exposures or banks or bed stabilized by concrete or other artificial materials are chosen for gaging sites where possible.

The simplest technique for measuring flow depth is to install a calibrated vertical scale along one bank and read depth from this scale. This technique was originally developed for use along the Nile River circa 620 AD. Most gages now use a stilling well in which the stage equals height of the water surface in the channel. A pressure sensor in the well is connected to instruments that record fluctuations in the stage through time, using something as simple as a paper strip chart and a stylus, or sending the electronically transduced measurements to a central receiving and recording station via satellite transmission. Discharge can be measured at regular intervals, the length of which typically reflects some tradeoff between accuracy and cost. Various statistical measures such as mean daily, mean monthly, and mean annual discharge can be computed from these data.

Supplemental Section 3.2.1 provides further information on gaging techniques and illustrations of different types of gages.

Direct discharge measurements are relatively limited. Some large rivers have continuous records extending back more than a century. More commonly, records are of short duration or with time gaps, and include only partial spatial coverage of a river network such that the main channel and a limited number of tributaries have one or more gaging stations, but many tributaries have few or no gages. Under these circumstances, a number of techniques can be used to indirectly estimate the discharge of a discrete event, such as a flood, or to estimate statistical characteristics of discharge through time, such as mean annual flow.

In regions with reasonably complete spatial and temporal coverage of discharge gages, the existing records can be composited to create *regional discharge–drainage area relations*. These relations can then be applied to estimate flows at ungaged sites based on the drainage area of the ungaged site (Sanborn and Bledsoe, 2006; http://streamstats.usgs.gov/index.html). Related to this is the practice of *storm transposition*, in which a particularly extreme storm for which records exist at one gage site is used to estimate the discharge that would result at another site if such a storm occurred (Changnon, 2002). This type of extrapolation can be inaccurate where discharge–drainage area relations are nonlinear as a result of changes in hydroclimatology with elevation or changes in rainfall–runoff relations with geology or land cover (Pitlick, 1994).

The most extreme precipitation input physically likely at a given site, the *probable maximum precipitation*, is defined as the greatest depth of precipitation for a given storm duration. This value is used to estimate the resulting *probable maximum flood*, which is mandated for design of structures such as dams where failure of the structure would result in loss of human life.

In the absence of direct measurements at discharge gages, flow can be indirectly estimated using several approaches. Most of these are based on a rearrangement of the *Manning equation* as presented earlier

$$Q = \frac{1}{n} A R^{\frac{2}{3}} S^{\frac{1}{2}} \tag{3.32}$$

where A is cross-sectional area, R is hydraulic radius, S is channel gradient, and n is the Manning roughness coefficient. The Manning equation is based on the assumption of steady, uniform flow so that slope, discharge, and velocity are constant with time and space along a segment of channel. This assumption may not apply very well to flow in natural channels, particularly during high flows. Nonetheless, the commonly used indirect methods of slope-area and step-backwater computations assume steady, uniform flow and use 1D hydraulic theory as their basis (Webb and Jarrett, 2002).

Both the slope-area and step-backwater methods use the conservation of mass, conservation of energy, and Manning equations to calculate flow. Slope-area uses known water-surface elevations to compute discharge (Dalrymple and Benson, 1967), whereas step-backwater uses discharge to compute stage (O'Connor and Webb, 1988). Both of these approaches, and use of the Manning equation, assume that a water-surface profile can be used with surveyed cross-sectional geometry to calculate R as an approximation of flow depth. The accuracy of this calculation depends on the representativeness of the surveyed cross-sectional geometry and water-surface profile. Geometry and water-surface elevation can vary along a river reach, and can vary through time, even during a single flood, because of scour and fill of the bed, rapid changes in flow, substantial sediment transport, and flow transitions between subcritical and supercritical (Jarrett, 1987; Sieben, 1997).

Variations of parameters through time are particularly problematic when the Manning equation is used to calculate flood discharge. The accuracy of indirect flood discharge estimations based on the Manning equation typically declines in at least four scenarios. The first scenario is rivers with very short duration and high magnitude discharges that do not have steady, uniform flow. This scenario is characteristic of smaller catchments, particularly in arid regions and the tropics. The second scenario is rivers with seasonal ice in which much of the runoff during the season when the ice is melting occurs on top of the ice (Priesnitz and Schunke, 2002). A third scenario that limits accuracy of indirect flood discharge estimations is on small, steep rivers with very rough boundaries (Bathurst, 1990). Finally, indirect discharge is difficult to estimate on rivers such as sand-bed channels that can undergo substantial changes in cross-sectional geometry during a flood as a result of erosion of the bed on the rising limb and deposition on the bed on the falling limb.

The Manning equation can also be used to estimate the magnitude of discharges that occurred in the distant past, if cross-sectional geometry and water-surface profiles are preserved. Information on water-surface elevation can come from *historical records* created by people, including physical marks on buildings or bedrock channel walls, or diaries, journals, damage or insurance reports (Gurnell et al., 2003).

Information on water-surface elevation can also come from botanical and paleostage indicators. *Botanical records* of flow stage take the form of damage to riparian trees caused by flood-borne debris that leaves a scar on the tree (Hupp and Bornette, 2003). *Paleostage indicators* in the form of erosional or depositional records of maximum stage can be used to reconstruct a water-surface profile at sites such as bedrock canyons where the cross-sectional geometry changes relatively slowly (Wohl and Enzel, 1995; Jarrett and England, 2002; Benito and O'Connor, 2013) (Table 3.3). Using each of these types of information, A, R, and S are directly measured from existing channel geometry and high-water marks, and n is estimated, allowing at least an approximate value of Q to be calculated.

Other indirect methods can be used to estimate unrecorded discharge. *Regime-based reconstruction* uses sedimentary features to reconstruct the cross-sectional geometry of relict channels preserved on a floodplain or in cut banks of the contemporary channel (Jacobson et al., 2003). Channel geometry is then related to a relatively frequent flow, such as mean annual flood, using an approach such as the Manning equation. *Competence estimates* use the average clast size, bedforms, or bedding structure to estimate past mean flow, or use the largest clasts likely to have been transported by fluvial processes to estimate the shear stress, stream power, or velocity of the associated peak flow (Wohl and Enzel, 1995). In addition to the scars mentioned earlier, several other characteristics of riparian vegetation can be used to infer past flow and channel dynamics (Table 3.3) (Yanosky and Jarrett, 2002).

Supplemental Section 3.2.1 provides more information on the use and relative accuracy of

Table 3.3 Types of paleoflood and paleoflow indicators.

Category	Types and information
Historical	Written records of date and extent and/or depth of flood
	Photographs of flood
	Maps of flood extent
	Physical marks recording peak stage
Botanical	Vegetation structured by age or type
	Impact scars record flood stage and date
	Adventitious stems or split-base sprouts record flood date
	Adventitious roots indicate burial by overbank sedimentation
	Exposed roots indicate date of bank erosion
	Variations in tree-ring width and symmetry indicate high and low flows
Geologic	Regime based: channel dimensions related to mean flow
	Competence: average clast size, bedforms, bedding structure related to mean flow; maximum clast size related to peak flow
	Paleohydraulic: depositional or erosional indicators of hydraulics (e.g., lateral gravel berms, berms at downstream end of plunge pools, lateral potholes)
	Stage indicators: scour lines, lichen limits, truncation of landforms impinging on channel (e.g., alluvial fans), silt lines, organic debris, boulder bars, slackwater sediments

different techniques for indirectly estimating ungaged discharge.

Different measures of discharge, from flood peak flow to average annual discharge, can also be numerically simulated. The use of existing discharge records to estimate flow at ungaged sites is based on statistical analyses and the assumption that represented trends are consistent and continuous and can therefore be extrapolated across space and time, as well as extrapolated to the very large flows that may not be present in the gage records.

An alternative to the statistical approach of estimating discharge is a deterministic approach based on a mechanistic concept used to simulate precipitation inputs and discharge outputs under imposed boundary conditions (Sieben, 1997). Deterministic approaches include a variety of models, such as the U.S. Army Corps of Engineers' HEC-1 model that estimates discharge resulting from rainfall.

Hydrologic models of rainfall–runoff processes at the watershed scale depend on input data. Input data can include topography of the catchment surface and parameters that are spatially distributed over (land cover, land use) and beneath (soil texture and moisture) the surface. Input data can also include the spatial characteristics of the river channel network (e.g., drainage density) and channel geometry. Many of these data are acquired using remote sensing and manipulated using GIS software. This scale of hydrologic modeling presents numerous challenges because catchment-scale models integrate data derived from a wide range of sources, at a range of spatial scales and resolution, and over a period of time (Downs and Priestnall, 2003). Understanding the sources and magnitude of uncertainty in the diverse input data is particularly important for evaluating the accuracy and representativeness of these models.

Data on the spatial distribution of basic hydrological variables such as rainfall and runoff can be combined with digital terrain analysis to estimate discharge at varying points in a channel network and to create spatially explicit water budgets (Montgomery et al., 1998). Marks and Bates (2000) provided an example of combining different types of models and data sets for modeling flood parameters. Working on the River Stour in the United Kingdom, they used a two-dimensional (2D) model of flow in the channel and across the floodplain with LiDAR-derived topographic data to simulate the extent of flood inundation, as well as local values of hydraulic parameters such as flow depth and velocity.

Another approach to numerically modeling floods involves combining gaged and ungaged streamflows with physical and statistical data to develop probability models for all uncertain parameters (Campbell, 2005). Hydrologists accept the fact that hydrologic model prediction is not deterministic, but rather must explicitly include an estimate of uncertainty. Uncertainty in model predictions reflects measurement errors for input and output parameters, model structural errors from the aggregation of spatially distributed real-world processes into a mathematical model, and errors of parameter estimation. Bayesian model selection uses the rules of probability theory to select among different possibilities, automatically choosing simpler, more constrained models. This approach can be applied to rainfall–runoff models (Sharma et al., 2005) or to models of streamflow, among other things. Both types of models can also be run using a Generalized Likelihood Uncertainty Estimation scheme or GLUE (Beven and Binley, 1992). The GLUE framework was one of the first attempts to represent prediction uncertainty. This framework conditions the parameter distributions and generates prediction uncertainty envelopes that incorporate parameter uncertainties (Wyatt and Franks, 2006). A great deal of attention is given to spatial variability in parameters and processes and to uncertainty in parameter estimation (Sivapalan et al., 2006).

Although direct measurements of flow stage at gaging stations include errors and uncertainty, these data remain the standard against which indirect estimates, statistical extrapolations, and numerical simulations are validated. Changing priorities in government expenditures in countries such as the United States, however, continue to result in declining numbers of active gages. Estimation of discharge using satellite-based remote imagery may help to fill this widening gap in discharge data, but such approaches are still being developed (Legleiter et al., 2009; Flener et al., 2012).

3.2.2 Flood frequency analysis

The frequency with which a flood of a given magnitude recurs is critical to understanding the physical and ecological significance of floods, as well as mitigating hazards associated with flooding. Flood-frequency analysis uses different techniques to relate flood magnitude to recurrence interval. The record of floods at a site can be considered in terms of an *annual maximum series*, which includes only the largest discharge in each year of record, or a *partial duration series* that includes all floods above a specified discharge. The partial duration series is likely to be more important geomorphically, particularly if the specified discharge has some physical relevance to sediment transport or channel morphology.

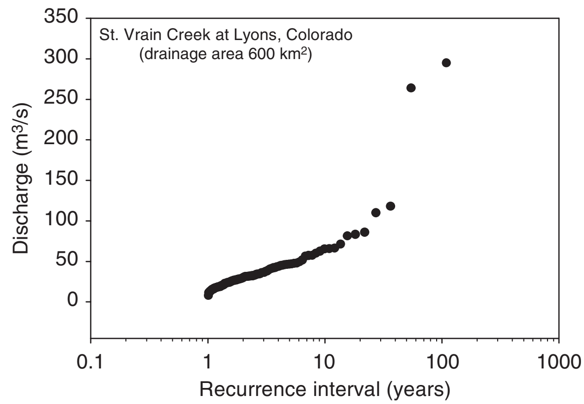

Figure 3.17 Sample flood–frequency curve: here, for St. Vrain Creek in Colorado, USA (drainage area 600 km^2). The typical annual peak flow results from snowmelt, but at much longer recurrence intervals, convective storms can generate very high magnitude flash floods. The two outliers in the upper right of this plot are rainfall flash floods that occurred in 1941 (largest flood) and 1919.

The simplest way to calculate recurrence interval T (years) for either annual maximum or partial duration series is via the Weibull formula

$$T = (n + 1)/N \quad (3.33)$$

where n is the number of years of record and N is the rank of a particular flood when floods are ranked from largest to smallest (Dalrymple, 1960). T can then be plotted against Q to create a *flood–frequency curve* (Figure 3.17). In any one year, the probability of the largest flow exceeding a flood with a recurrence interval of 10 years is 1/10. A flood with an average recurrence interval of 10 years is commonly referred to as a 10-year flood, but this type of abbreviated description can be very misleading. There is no physical reason that a 100-year flood cannot occur during two consecutive years; it is improbable, but not impossible.

There are no strong theoretical justifications for applying any particular statistical distribution to hydrologic data. Log-Pearson Type III is commonly used in the United States because it is a skew distribution bounded on the left and therefore of the general shape of most hydrologic distributions. The recommended procedure for using the log-Pearson distribution is to convert the data to logarithms and compute the value of a flood of specified probability, X, for any probability level as

$$\log X = \overline{\log X} + K\sigma_x \tag{3.34}$$

where $\overline{X}$ is the mean of the flood series, K is obtained from a standard table, and σ_x is the standard deviation of the series. The Gumbel extreme value distribution is commonly used in the United Kingdom.

Regardless of the distribution used, treating the highest known discharges for a site is particularly problematic because these discharges tend to be outliers that do not follow trends present in the remainder of the data. Historical and paleoflood information can be combined with systematic gage measurements of discharge to extend the length of the flow record at a site. The historical and paleoflood data must be treated differently than the systematic data because they represent what is known as a *censored record*: only floods above a magnitude threshold are recorded, rather than all flows being recorded, as at a stream gage. Early work used threshold-exceedance maximum likelihood estimators, which employed the number of floods that exceeded a known threshold, without differentiating the magnitude of each flood (Stedinger and Baker, 1987). Subsequent work relies on moments-based parameter estimation procedures (Cohn et al., 2001; England et al., 2003) or simply uses field evidence of paleoflood magnitude to quantify nonexceedance bounds for discharge over a time interval defined using various geochronologic techniques (England et al., 2010).

Another difficult issue for flood-frequency analysis is the existence of *mixed distributions*. Many sites include floods produced by more than one hydroclimatic mechanism—for example, flash floods resulting from convective thunderstorms and snowmelt floods, or flash floods and longer duration floods caused by dissipating hurricanes. Analyses that separate populations with different flood-generating mechanisms result in improved parameter estimates of the component distributions and a better understanding of the physical basis for extreme floods (Hirschboeck, 1987).

The presence of nonstationarity is a primary concern for any form of flood-frequency analysis. *Stationarity* is the assumption that any statistical property of the flow record, such as mean annual flow or mean annual flood peak, has a probability density function that does not vary through time. This assumption facilitates extrapolation of trends in hydrologic data forward in time and prediction of the magnitude of future 100-year floods, for example, based on the magnitude of 100-year floods in the past, or in a situation where the record length is less than 100 years.

Assumptions of stationarity can be incorrect if the particular period of record represents some deviation from longer-term averages. Precipitation can exhibit long-period variability such that more zonal or meridional circulation patterns characterize periods up to several decades in length at regional to continental scales (Hirschboeck, 1988). More meridional circulation tends to produce more severe flooding, so using 20 years of record from a period of meridional circulation to predict future flood frequency might overestimate the recurrence interval of a particular flood magnitude. Other climatic circulation patterns can produce fluctuations in flooding at timescales of decades (Webb and Betancourt, 1990) to centuries and millennia (Ely et al., 1993; Benito et al., 1996; Redmond et al., 2002) across broad regions. Human alterations of rainfall–runoff relations as a result of changing land cover or flow conveyance in river networks can also compromise the assumption of stationarity. Finally, ongoing global climate change associated with CO_2-induced warming is rendering the assumption of stationarity inappropriate in many regions. Efforts are underway to develop nonstationarity probabilistic models (Milly et al., 2008), but no widely used method is yet available.

3.2.3 Hydrographs

A *hydrograph* is a curve of discharge plotted against time. An *event* or *flood hydrograph* represents a single flood. An *annual hydrograph* represents flow over the course of an average year as derived from averages of some flow interval (typically, minutes to 1 day) over the period of gage records. A *unit hydrograph* is based on runoff volume adjusted to a unit value (e.g., 100 mm of precipitation spread evenly over the drainage basin) (Rodriguez-Iturbe and Valdes, 1979). A *geomorphologic unit hydrograph* relates the unit hydrograph to the morphologic parameters of the river network and defines the travel time distribution of water particles to the outlet of the basin (Rodriguez-Iturbe and Valdes, 1979).

Any hydrograph includes base flow and direct runoff, a rising limb, a peak, and a falling limb (Figure 3.18). *Base flow* results primarily from groundwater inputs to a channel and is the stable, low

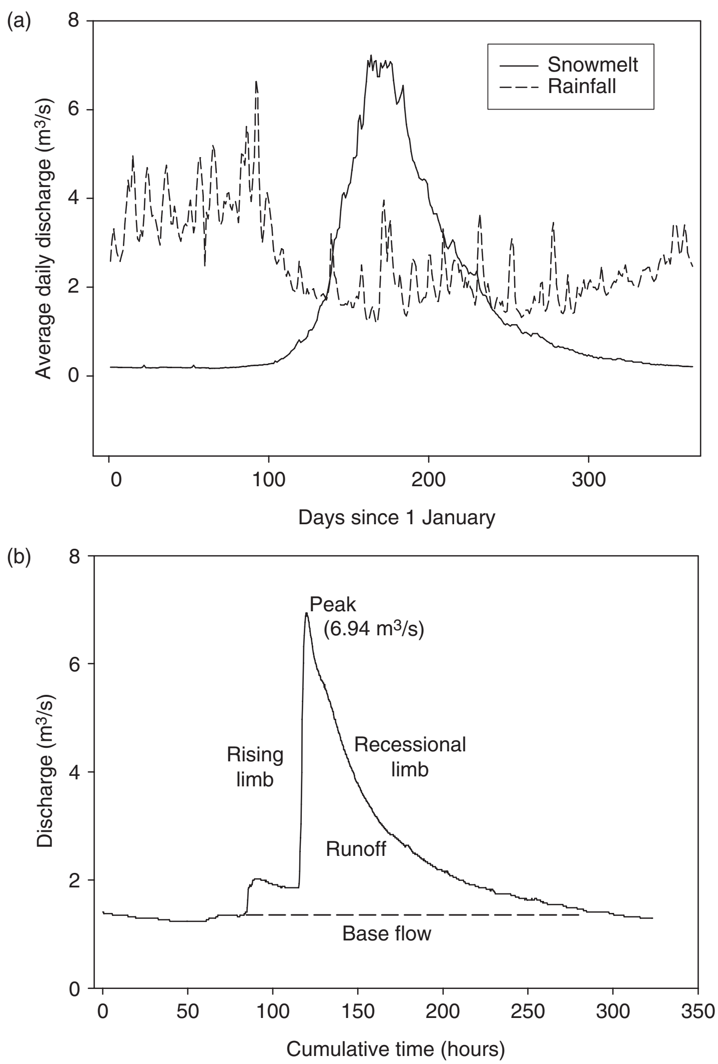

Figure 3.18 Sample hydrographs. (a) Average annual hydrographs based on average daily discharge over the period of record for a snowmelt-dominated stream (North St. Vrain Creek, Colorado, USA; drainage area 91 km^2) and a rainfall-dominated stream (Cedar Creek, South Carolina, USA; drainage area 161 km^2). (b) Flood hydrograph for a single rainfall-generated flood on Gills Creek, South Carolina; drainage area 167 km^2).

flow to which a river returns following precipitation inputs. *Direct runoff* results from the combined inputs of overland flow and flow in the unsaturated zone (Section 2.2.1). The *rising limb* is the portion of the hydrograph where discharge increases as a result of direct runoff inputs, the *peak* represents the maximum discharge within a particular timespan, and the *falling or recession limb* reflects progressive declines in direct runoff with time until discharge returns to base flow.

The term *flashy* is commonly applied to hydrographs with short duration peaks and rapid rise and recession. Baker et al. (2004) proposed a flashiness index

$$\text{Richards–Baker Flashiness Index} = \frac{\sum_{i=1}^{n} |q_i - q_{i-1}|}{\sum_{\mathrm{i}=1}^{n} q_{\mathrm{i}}} \tag{3.35}$$

where q is mean daily flow, q_{i} is flow on a particular, and $\mathrm{q}_{\mathrm{i}-1}$ is flow on the following day. Although many drainages are described as being flashy, the criteria used for this designation vary widely and are not necessarily comparable between studies.

The *lag time* or *basin lag* indicates the delay between the center of mass of precipitation inputs and the center of mass of streamflow. Lag time reflects the manner in which precipitation is transmitted from hillslopes into the channel, as well as the size of the catchment, spatial arrangement of channels in the river network, and geometry of valleys and channels. Lag time for a given type of precipitation input increases with basin size because direct flow must travel longer distances along hillslope paths before reaching a channel, and longer distances within the river network before reaching the basin outlet. Lag time is lower for equant-shaped basins than for linear basins. Equant-shaped basins tend to produce larger, shorter peak discharges because of efficient concentration at the basin outlet (Strahler, 1964). Lag time also increases where broad valley bottoms and floodplains allow direct runoff to move relatively slowly downstream in overbank areas or multiple channels.

Hydrograph characteristics can also be strongly influenced by the *hydroclimatology*, or precipitation regime. Ignoring other influences such as drainage basin size, snowmelt runoff tends to be spread over a longer time and to produce a peak flow of longer duration and lower peak than rainfall runoff. Different types of rainfall produce differently shaped hydrographs. Convective rainfall, because of its greater intensity and short duration, tends to produce a more peaked hydrograph than cyclonic rainfall. Rain-on-snow, as the name implies, occurs when rain falls on a snowpack. The resulting runoff depends on the magnitude of the rain and the water equivalent and spatial extent of the snowpack, and can be quite large but of shorter duration than snowmelt (McCabe et al., 2007). Because of these characteristic differences in event and annual hydrographs, a river's flow regime is typically described as being *snowmelt dominated* or *rainfall dominated*. These categories are not mutually exclusive because different types of flooding can occur during different seasons or over a period of many years in the same region. For example, a snowmelt-dominated hydrograph can have secondary peaks associated with late summer convective storms, or a rainfall-dominated hydrograph can have predominantly frontal cyclonic precipitation with occasional hurricane rainfall.

In addition to describing flow regime in terms of the dominant type of runoff, flow regime can be characterized by its spatial and temporal continuity. *Perennial flow* is continuous through time and space: some level of flow is always present throughout the river network so designated. *Intermittent flow* is spatially discontinuous such that some portions of a river contain flow while other portions are dry (Figure S3.20). This situation can arise where a channel crosses into a different climate, as when perennial river segments flowing from mountainous highlands with large annual precipitation enter drier lowlands and the surface flow evaporates or infiltrates. Intermittent flow can also result from longitudinal contrasts in subsurface permeability, as when flow in a channel crossing a thick, permeable alluvial layer infiltrates and moves downstream in the subsurface, returning to the surface channel where an impermeable layer close to the surface forces the water upward once more. *Ephemeral flow* is temporally discontinuous, with periods of surface flow shortly after precipitation inputs to the basin interspersed with periods of no flow.

Intermittent and ephemeral flows are most likely to occur in dry environments such as arid and semiarid regions of the polar, temperate, and tropical latitudes, and in very small catchments that may have limited base flow inputs. Intermittent flow can also occur in *karst terrains* where underground drainage networks developed through chemical dissolution of carbonate rocks capture surface flow in some portions of a network. Surface–subsurface exchanges in karst environments can produce *blind valleys* that end suddenly where a stream disappears underground and *exsurgence* where an underground stream with no surface headwaters reaches the surface.

The spatial and temporal extent and magnitude of flow also influence the connectivity of river networks. Ecologists distinguish *riverine connectivity*, which indicates spatial linkages within rivers, and *hydrologic connectivity*, which is the water-mediated transport of matter, energy, and organisms within or between elements of the hydrologic cycle (Freeman et al., 2007). Both forms of connectivity are vital to maintaining the *ecological integrity* of riverine ecosystems, where ecological integrity is the undiminished ability of an ecosystem to continue its natural path of evolution, its normal transition over time, and its successional recovery from perturbations (Westra et al., 2000).

The typical scenario is that discharge and occurrence of perennial flow increase downstream within a river network as contributing drainage area increases. If the rate of increase in discharge with drainage area is known, the more easily measured drainage area can be used as a surrogate for discharge. This may not be the case, however, in drylands (semiarid and arid regions). Low annual precipitation and high evaporation in drylands result in low annual runoff, and the interannual variability of runoff increases in the driest regions. Spatial variability of precipitation and runoff limit the use of drainage area as a surrogate for discharge in drylands (Tooth, 2013). Instead, most dryland river networks exhibit downstream decreases in discharge during floods as a result of transmission losses from infiltration into unconsolidated alluvial channel beds, and evaporation and transpiration, as well as a common absence of appreciable tributary inflows in the lower parts of many dryland river networks (Tooth, 2013). Maximum flood peaks and discharge per unit drainage area can be quite large in dryland rivers, particularly where small, steep catchments and abundant low-permeability surfaces associated with bedrock or crusted soils are present. This situation can result in highly skewed flood–frequency distributions and steep flood–frequency curves that reflect a high ratio of large to small flows (Knighton and Nanson, 2002; Tooth, 2013).

3.2.4 Other parameters used to characterize discharge

Discharge measurements that are continuous during a year and over several years can be used to establish a *flow–duration curve* in which mean discharges over a specified time interval (e.g., daily mean) are grouped into selected classes of discharge magnitude and plotted against the percentage of time that class is equaled or exceeded (Figure 3.19). The shape of the flow–duration curve reflects the drainage basin's response to precipitation inputs. The steeper the flow–duration curve, the more quickly storm runoff enters the channel.

At sites with measurements of suspended sediment concentration at varying discharges that allow construction of a *sediment rating curve*, the flow–duration curve can be used to develop a *cumulative sediment transport curve* indicating how flows of differing magnitude and recurrence interval contribute to sediment transport during an average year.

Ecologists developed the *flood pulse concept* for large, lowland rivers in recognition of the importance of the magnitude, timing, duration, and rates of rise and fall of annual or seasonal floods to aquatic and riparian ecosystems. Natural, predictable floods—the flood pulse—that inundate at least a portion of the floodplain are associated with much higher biological productivity. The flood pulse provides clear, shallow water in overbank areas

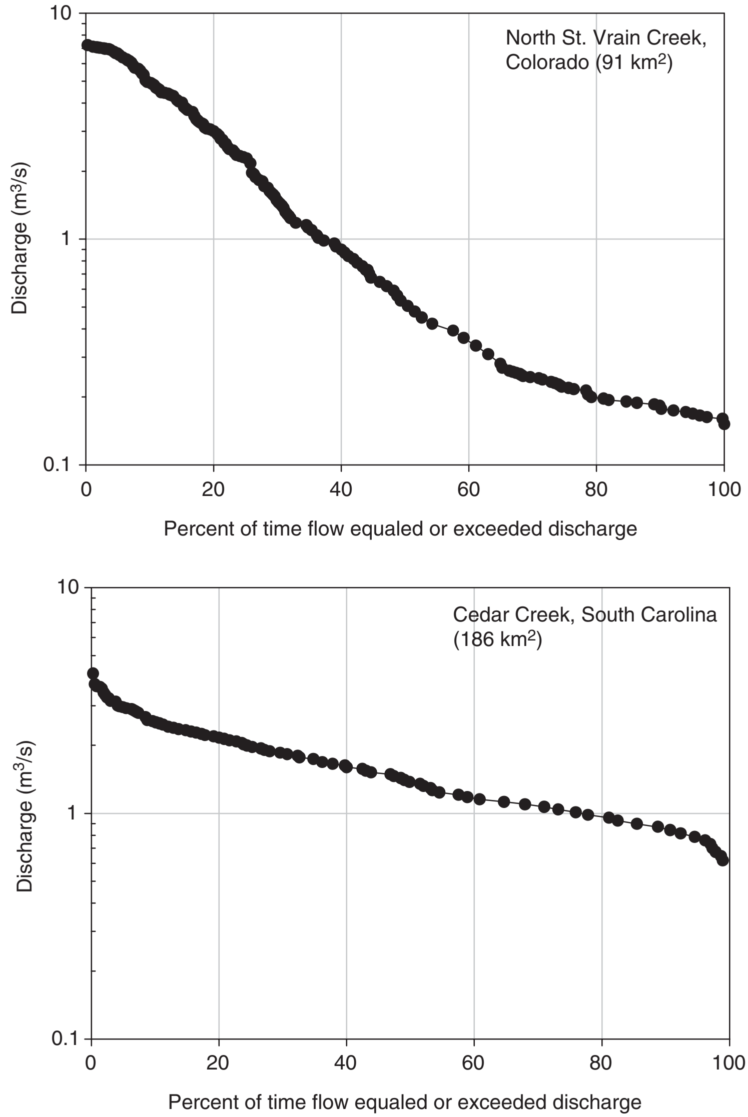

Figure 3.19 Sample flow–duration curves for North St. Vrain Creek, Colorado (snowmelt flow regime) and Cedar Creek, South Carolina (rainfall flow regime).

for primary, photosynthetic production. The flood pulse also creates fish nursery habitat and feeding, as well as more abundant and diverse habitat for a variety of other organisms. Inundation of overbank areas and then subsequent recession of streamflow into channels during the flood pulse facilitates the exchange of nutrients, organic matter, and organisms between the channel and floodplain (Junk et al., 1989; Bayley, 1991).

Ecologists also describe *flow pulses* that are fluctuations in surface waters below the bankfull level of a channel (Tockner et al., 2000). Although these fluctuations do not cause overbank flow, they change the extent of flow in secondary channels and in areas of flow separation along a single, confined channel and provide some of the same biogeochemical exchanges and habitat abundance and diversity as overbank flows.

A third key concept in riverine ecology is the *natural flow regime* (Poff et al., 1997). This phrase refers to the magnitude, frequency, duration, timing, and rate of change of flow conditions in the absence of human interference through processes such as flow regulation. Channel geometry and aquatic and riparian communities are typically adjusted to the natural flow regime so that altering this flow regime is likely to cause corresponding alterations in the physical and ecological characteristics of the river. Management of rivers with flow regulation increasingly emphasizes protecting or restoring at least some approximation of the natural flow regime, as discussed in Chapter 7.

3.2.5 Hyporheic exchange and hydrology

The hyporheic zone is the portion of unconfined, near-stream aquifers where stream water is present. Hydrologists define this zone as a flow-through subsurface region containing flow paths that originate and terminate at the stream. Gooseff (2010) proposed that the hyporheic zone be defined based on the timescale of flow, analogous to definitions of floodplains based on inundation frequency, such that there are 2–24-hour hyporheic zones. The hyporheic zone can extend up to 2 km laterally from the active channel along rivers with broad, gravel floodplains (Stanford and Ward, 1988) (Figure S3.21). Hyporheic flow may constitute only a minor portion of stream discharge (<1%) in steep, headwater channels with limited alluvium (Wondzell and Swanson, 1996), but can be 15% or more of surface discharge in larger, lowland alluvial rivers (Laenen and Risley, 1997). In the latter case, ignoring hyporheic flow can lead to errors in discharge estimation, particularly given that the characteristics of the hyporheic zone vary widely in space and time.

Supplemental Section 3.2.5 provides more information about the processes of, and controls on, hyporheic exchange.

3.2.6 River hydrology in cold regions

Hyporheic and groundwater exchange can certainly complicate the relations between precipitation inputs and outputs of river flow, but in most basins, greater precipitation inputs equate to higher river stage and greater discharge. This is not necessarily the case in high-latitude rivers with strongly developed seasonal ice cover. A river-ice season that lasts more than 100 days between autumn freeze-over and spring breakup characterizes many rivers of high latitudes, including some high-elevation or interior rivers as far south as 42° N in North America and 30° N in Eurasia (Prowse and Beltaos, 2002). The hydraulic resistance added by an ice cover elevates river water levels, particularly when the ice cover is hydraulically rough during freeze-over and breakup.

River ice can modify the quantity of flow in a channel via three mechanisms (Prowse and Beltaos, 2002; Ettema and Kempema, 2012). The first involves reduction of contributing area, when anchor ice frozen to the river bed cuts off groundwater inflow. The second involves direct storage of water in river ice, which typically involves slower abstraction during freeze-over and rapid re-supply during breakup. The third mechanism, which can produce the most significant effect, is hydraulic storage in the channel. This involves increased water levels because of the increased hydraulic resistance caused by an ice cover. The latter effect can be particularly significant when ice jams form during breakup.

Ice-jam frequency and the resulting floods are highly variable from year to year (Boucher et al., 2012) as a result of the strong influence of interannual fluctuations in air temperature. The reduction in flow created by the three mechanisms described earlier can be equivalent to nearly 30% of the flow that would otherwise occur at a particular site (Prowse and Carter, 2002). Release of this water during breakup can account for nearly 20% of the spring peak flow (Prowse and Carter, 2002). Because of these effects, early winter freeze-over can create the lowest discharge of the year, even though runoff is lowest during late winter, and breakup frequently establishes the annual maximum water levels, even though maximum discharge is more likely to result from spring snowmelt or summer rainfall later in the year (Prowse and Ferrick, 2002). One implication of the presence of ice is that drainages with ice cover may not exhibit the direct relationship between flow and stage that exists in most rivers (Prowse and Beltaos, 2002).

Flow regimes in high-latitude rivers can be distinguished as nival, proglacial, wetland, and prolacustrine (Woo and Thorne, 2003). Snowmelt and river ice breakup generate the largest flows in *nival regimes*. High flows are prolonged into the summer by glacial melt water in *proglacial regimes*; some of the highest flows occur in mid to late summer when glacier melt is most pronounced. Rivers below wetlands also have prominent snowmelt peaks because the wetlands have low storage capacity while frozen, but once the ground thaws, the wetlands can retain water and retard summer flows in *wetland regimes*. The influence of large lakes creates a fairly even runoff throughout the year in *prolacustrine regimes*, although the outflow channel of an arctic lake is more likely to be blocked during freeze-over and inflow to the lake can be very small, minimizing winter outflow.

3.2.7 Human influences on hydrology

Human activities can alter flow regimes in channels indirectly through changes in climate, topography, and land cover that influence the amount and downslope pathways of water entering channels from uplands (Supplemental Section 2.2.1). Direct human alterations of streamflow involve changes in the volume and timing of flow as a result of flow regulation, and changes in the ability of channels to convey flow downstream. Flow regulation includes dams and diversions.

Dams can be small barrages or weirs that are frequently overtopped during moderate to high discharges, but serve to pond water upstream during low discharges. At the opposite end of the size scale, large dams are those taller than 15 m and major dams are taller than 150 m. More than 45,000 large dams and more than 300 major dams currently exist. More than half (172 of 292) of the world's large river systems are affected by dams (Nilsson et al., 2005). Only about 2% of the total river kilometers in the United States are not affected by dams (Graf, 2001), and other industrialized countries have equally high rates of flow alteration by dams.

Dams vary in their design and operation. Dams that store water for flood control or water supply, for example, release the stored water in large, sustained pulses. Dams designed for hydroelectric power generation typically have substantial daily fluctuations in water releases to maximize power production during periods of high demand. The typical effect of any type of dam, however, is to homogenize regional river flow regimes (Poff et al., 2007) by reducing peak flows and reducing the variability in lower flows.

Flow diversion can either reduce flow in the source stream or augment flow in the receiving stream. In extreme cases, all water is removed from the source stream during at least a portion of the year, or the peak flow is more than doubled in the receiving stream. Although flow diversion has historically been most common on small- to medium-size rivers, massive diversion projects on very large rivers are now underway or being planned in China and parts of Africa and the Middle East.

Direct alterations of channel form that change flow conveyance and at least some characteristics of the hydrograph take several forms. Construction of levees prevents or limits overbank flows, increasing the magnitude and velocity of water conveyed within the channel during higher discharges. Channel engineering in the form of bank stabilization, dredging, or straightening (channelization) typically increases cross-sectional areas, reduces boundary resistance, and limits exchanges between sediment in transport and sediment in the channel bed and banks. All of these alterations typically result in larger magnitude but shorter duration flood peaks. Removal of beavers (*Castor fiber* in Europe, *Castor canadensis* in North America) and their small dams typically reduces overbank flooding and associated floodplain wetlands, and increases flood peak magnitude and velocity by confining flood discharges to a main channel. Alteration of riparian vegetation occurs either through removal of vegetation or through introduction of exotic species with substantially different characteristics than the native species. These alterations change bank

Table 3.4 Types of direct human influences on streamflow.

Type of influence	Effects	Sample references
Flow regulation	Dams typically reduce the mean and coefficient of variation of annual peak flow, increase minimum flow, shift the seasonal variability, and increase diurnal fluctuations; diversions typically reduce base flows and flood peaks in the source stream and increase base and peak flows in the receiving stream	Graf (2006); Ryan (1997), Wohl and Dust (2012)
Channel engineering	Levees prevent or limit overbank flow and thus typically increase the magnitude and decrease the duration of peak flows	Kondolf (2001)
	Bank stabilization can decrease channel complexity and increase conveyance	
	Channelization (straightening and dredging) increases channel conveyance and typically increases the flashiness of streamflow	Shankman and Pugh (1992)
Alterations of riverine biota	Removal of riparian vegetation decreases near-bank and overbank roughness, can decrease bank stability, decreases transpiration	Anderson et al. (2006a)
	Introduction of exotic riparian vegetation can change vegetation density and function, typically with increased density and roughness that can facilitate channel narrowing	Dean and Schmidt (2011)
	Removal of beaver decreases channel–floodplain connectivity, increases flow velocity and peak flow magnitude, decreases elevation of alluvial water table	Wohl (2001)

stability, near-bank and overbank roughness, and transpiration rates (Wohl, 2000a). Table 3.4 summarizes the effects of different types of direct human influences on streamflow.

Human effects on nearly all aspects of river form and function are now nearly ubiquitous. Sediment fluxes are very difficult to quantify accurately as a result of spatial and temporal fluctuations within each drainage basin, as well as incomplete systematic measurements of sediment flux in most basins, but global-scale estimates have been published. People have simultaneously increased the sediment transport by global rivers through soil erosion (Hooke, 2000) by more than 2 billion metric tons per year (of a total estimated 14 billion tons per year prior to human alteration), but reduced the flux of sediment reaching the world's coasts by 1.4 billion metric tons per year because of retention within reservoirs behind dams (Syvitski et al., 2005). Global manipulations result in a doubling to tripling of the residence time of continental runoff and a 600%–700% increase in fresh water stored in channels within reservoirs (Vörösmarty et al., 2004).

Our activities alter the water reaching Earth's surface, alter how that water moves downslope and into channels, and alter how the water moves downstream through river networks. These cumulative alterations are extremely intensive and widespread globally.

Supplemental Section 3.2.7 provides more details and example case studies of indirect and direct human alterations of streamflow.

3.3 Summary

Equations used in basic engineering hydraulics start from assumptions that flow in a channel is uniform and steady, and that the channel has relatively smooth boundaries and limited turbulence. Natural channels seldom meet these assumptions, and the degree to which flow is varied and unsteady determines how far actual hydraulic characteristics deviate from expected conditions. The instabilities associated with hydraulically rough boundaries, flow transitions, and spatial and temporal variations in velocity and turbulence are inherent in natural channels, as are the interactions between flowing water and the channel boundaries that result in sediment movement and channel adjustment. A substantial amount of research is focused on quantifying and predicting hydraulic variables such as resistance (*n* or *f*), velocity, turbulence intensity, shear stress, and stream power because these variables can be used to predict sediment movement and channel geometry. Our ability to accurately characterize hydraulics in natural channels is greatest in those channels that best approximate the assumptions underlying engineering hydraulics—sand-bed channels with readily deformable boundaries and simple geometry. Natural channels with more hydraulically rough boundaries, more complex and spatially variable geometry, and greater erosional thresholds—gravel-bed, boulder-bed, and bedrock channels—are typically quite poorly characterized using standard assumptions from engineering hydraulics.

The energy of flow quantified in hydraulic equations inherently reflects the volume of water moving down a channel, so a substantial amount of research also focuses on quantifying and predicting discharge. Systematic, gage-based measurements of discharge seldom provide sufficient length and spatial density of record to infer the characteristics of very large, infrequent flows. A variety of indirect methods are used to estimate the magnitude and recurrence interval of a wide range of flows, from rare floods to the shape of an average annual hydrograph. Human activities have directly and indirectly altered river discharge for centuries in much of the world, making it challenging to infer a natural flow regime in the absence of human manipulation.

Channel processes II

Chapter 4
Fluvial sediment dynamics

This chapter examines the movement of particulate and dissolved sediment within river channels. The chapter is divided into four main sections. The first section describes the channel bed and initiation of sediment motion from the bed. Within this part of the chapter, the first subsection addresses a stationary streambed and the methods used to characterize bed sediment. The ability to characterize the bed substrate is important to understanding one source of sediment supply when sediment is in transport, as well as sources of flow resistance that influence the distribution of hydraulic forces moving sediment. The next two subsections discuss the initiation of motion of individual particles on a non-cohesive bed, and the removal of aggregates of particles from a cohesive bed.

The second section of the chapter focuses on the processes of sediment transport and deposition. Sediment can be transported downstream in solution, as particles suspended in the flow, or as particles that remain in contact with the bed. Each mode of transport is influenced by similar hydraulic forces, but is described by a distinct suite of processes and equations. Sediment moving in contact with the bed is differentiated between readily mobile and infrequently mobile bedforms. This division recognizes a fundamental distinction among rivers. Some rivers, particularly sand-bed channels, have relatively *mobile beds* in which sediment moves frequently and sometimes at relatively high rates, at least in terms of grain numbers. These channels are also referred to as live-bed or regime channels. In contrast, rivers with a bed dominated by gravel and larger particles are *threshold channels* in which bed sediment moves relatively rarely and at low rates. This section of the chapter ends with a description of bedforms in cohesive sediment, and processes and forms of sediment deposition within the channel boundaries.

The third section of the chapter discusses factors that control bank stability and processes of bank erosion. Bank erosion gets a great deal of attention because of the potential for property loss and damage to infrastructure as a result of bank failure. Erosion of banks is distinct from that of streambeds for at least two reasons. First, bank erosion commonly occurs via mass failures, as well as grain-by-grain detachment from the banks. Second, bank stability and erosion can be strongly influenced by vegetation.

The final section examines sediment budgets, which use information on sediment input and storage to quantitatively estimate fluxes of sediment past a given point in a watershed. Sediment yield is

Rivers in the Landscape: Science and Management, First Edition. Ellen Wohl.

Companion Website: www.wiley.com/go/wohl/riverslandscape

characterized by great spatial and temporal variability, with much of the sediment coming from a limited portion of a drainage basin during a relatively short period of time. Globally, a disproportionate amount of sediment originates in small, mountainous drainages.

4.1 The channel bed and initiation of motion

4.1.1 Bed sediment characterization

Knowledge of the size distribution, packing, sorting, shape of particles, and stratigraphy of bed sediments is important for understanding the initiation of motion of particles on the bed, sediment transport, and channel change. Accurately measuring the characteristics of the bed sediments becomes more difficult as these sediments become more coarse grained and heterogeneous, not least because progressively larger samples are necessary to accurately characterize the grain-size distribution. Consequently, a great deal of effort has been devoted to measuring the surface and subsurface bed sediments in gravel-bed rivers.

The grain-size distribution of bed sediments can be measured by taking a bulk sample and sieving the sediments in a laboratory. This works well for pebble-size or finer sediments, but quickly becomes impractical with coarser sediments because of the large sample size needed. As the intermediate axis of the largest clast approaches 64 mm, sample sizes required to accurately characterize the distribution exceed 400 kg (Church et al., 1987). Samples in which the weight of the largest clast is <1% of the total weight can provide unbiased estimates of mean grain size (Mosley and Tinsdale, 1985), although this makes sampling coarse sediments almost entirely impractical. *In situ* measurements of clast size can be more practical in terms of avoiding sediment collection, but such methods may not consistently sample clasts <15 mm in diameter (Fripp and Diplas, 1993). Volumetric samples of finer sediment can also be combined with *in situ* measurements of coarser sediment in a bed with mixed grain sizes (Bunte and Abt, 2001).

In situ methods for characterizing bed sediment can employ direct measurements of clast size along a grid or a random walk, as originally proposed by Wolman (1954). *In situ* approaches can also employ systematic subsampling, in which all particles exposed at the surface within a defined area are measured, or are removed from an area of the bed using an adhesive such as clay or wax: removal only works where particles are small enough to be lifted from the bed (and subsequently sieved) using an adhesive layer. *In situ* methods in medium to large rivers increasingly use Fourier or other spectral analysis of particle outlines on digitized photographs (Supplemental Section 4.1.1).

Much effort has gone into determining the minimum sample size necessary to accurately characterize bed sediment using either bulk or *in situ* sampling. The default method remains the 100 clasts originally proposed by Wolman (1954), which is usually sufficient for measures of the central tendency such as mean grain size. Fripp and Diplas (1993), Rice and Church (1996), and Bunte and Abt (2001) described how to determine the minimum sample size based on the size fraction of interest and the level of precision required. The desired precision depends on the intended use of the grain-size distribution, but commonly results in a sample size closer to 300 clasts. Measurements repeated at a site through time with the intent of monitoring the percentage of finer sediment on the surface of a gravel-bed stream as an indicator of fish or macroinvertebrate habitat, for example, may require greater precision than a reach-averaged estimate of bed roughness based on grain size.

Once the grain-size data are obtained, they can be used to develop a cumulative frequency distribution of frequency by weight or frequency by number (Supplemental Section 4.1.1). Either type of data can then be used to estimate D_x, where x is the percentage of the grain-size distribution that is finer than the reference size (e.g., D_{50} is the grain size for which 50% of the distribution is finer).

In addition to size distribution, bed sediment can be characterized in terms of density, shape, sorting, and packing. *Grain density* is the mass per unit volume of the sediment, which is usually assumed to be 2.65 g/cm^3—the density of quartz. (Bulk density of aggregates of grains is more typically in the range of 1.4–1.8 g/cm^3.) Grain density affects the fall or settling velocity, the velocity at which a grain moves down a water column. *Specific gravity* is the density of the sediment relative to the density of water (2.65 g cm^{-3}/L g/cm^3 = 2.65).

Grain shape includes form and surface texture of a sediment grain. Shape affects fall velocity: flatter particles fall more slowly than spherical particles of similar weight and density. Shapes can be described as spheres, blades, discs, and rods, and with respect to roundness and sphericity (Folk, 1980). Shape factors such as grain elongation (the ratio of the long axis to the intermediate axis) can also be quantified (Robert, 2003). The shape of a grain influences the surface area exposed to the flow and the ease with which the grain can be rolled along the bed.

Sorting describes the range of particle sizes present. Well-sorted sediments have a narrow range of particle sizes. Various measures of sorting exist, including the widely used inclusive graphic standard deviation (Folk, 1980):

$$\sigma_I = \frac{\phi 84 - \phi 16}{4} + \frac{\phi 95 - \phi 5}{6.6} \tag{4.1}$$

where $\phi\ x$ is the phi size for which x% of the distribution is finer grained: $\phi = -log_2 x$, where x is grain size in millimeter. Values of $\sigma_I < 0.35\phi$ are very well sorted, 0.35–0.5ϕ are well sorted, 0.5–0.71ϕ moderately well sorted, 0.71–1ϕ moderately sorted, 1–2ϕ poorly sorted, 2–4ϕ very poorly sorted, and >4ϕ extremely poorly sorted.

Packing describes the organization of particles on the bed. Packing can be qualitatively categorized as imbricated, interlocked, and open (Figure S4.3). *Imbricated* clasts lie with their long axes parallel to flow and dipping slightly up-current. Imbrication makes it more difficult to entrain clasts. Imbricated clasts can provide useful paleocurrent indicators when preserved in stratigraphy. Individual coarse particles touch one another in an *interlocked bed*, which is *framework supported*, whereas coarse particles are surrounded by finer sediment in a *matrix-supported bed*. *Open-framework gravels* lack finer sediment in the pore spaces between individual gravel particles. The distinction between framework supported and matrix supported can be important in contexts such as aquatic habitat. A framework-supported or open-framework bed allows water, dissolved oxygen, and nutrients to circulate more readily around fish eggs laid in the streambed, for example, than does a matrix-supported bed.

Additional information about characterizing bed sediment is included in Supplemental Section 4.1.1.

4.1.2 Entrainment of non-cohesive sediment

Most grains of sand size and coarser (Table 4.1) on a streambed are non-cohesive. *Entrainment* refers to the initiation of motion of sediment. Similarly, *incipient motion* describes the threshold condition when the hydrodynamic moment of forces acting on a particle balances the resisting moment of force (Shields, 1936; Julien, 1998). Being able to predict when sediment will be entrained is critical to understanding bed erosion, movement of bedforms, maintenance

Table 4.1 Grain-size categories.

Category	**Size range (mm)**	**Phi units[a] (ϕ)**
Boulder	≥256	−8 to −12
Cobble	64–256	−6 to −8
Gravel[b]	2–64	−1 to −6
Sand[c]	0.062–2	4 to −1
Silt	0.004–0.062	8–4
Clay	≤0.004	≤8

[a] Phi units are calculated from $\phi = -log_2 x$, where x is grain size in mm.
[b] Gravel is sometimes divided into pebble (4–64 mm, −2 to −6 phi) and granule (2–4 mm, −1 to −2 phi).
[c] Sand is sometimes divided into very coarse sand (1–2 mm, 0 to −1 phi), coarse sand (0.5–1 mm, 1–0 phi), medium sand (0.25–0.50 mm, 2–1 phi), fine sand (0.125–0.25 mm, 3–2 phi), and very fine sand (0.0625–0.125 mm, 4–3 phi).

and stability of aquatic habitat, sediment transport, and hydraulic roughness.

Studies of entrainment initially focused on well-sorted, sand-bed channels, and these remain the channels for which entrainment can be predicted most accurately. As cohesion between particles increases in grains finer than sand, or in bedrock, bed sediment is less likely to detach as individual particles. Under these conditions, substrate characteristics such as porosity and permeability or bedrock jointing influence entrainment. As the grain-size distribution becomes wider in coarse-grained channels, effects such as packing, sorting, shielding of finer particles, protrusion of coarser particles, and particle shape influence entrainment.

Modern studies of entrainment in sand-bed channels start with Shields (1936), who conducted flume experiments on the initiation of motion in channels with relatively uniformly sized sand grains. Shields quantified entrainment using *dimensionless critical shear stress*, τ_c^*, calculated by dividing the critical shear stress by an approximation of the weight of an immersed grain,

$$\tau_c^* = \frac{\tau_c}{(\rho_s - \rho)gD} \tag{4.2}$$

where τ_c^* is the dimensionless critical shear stress, ρ_s is the sediment density, ρ is the water density, g is the acceleration due to gravity, and D is the grain size. D_{50} is typically used for D when the grain-size distribution is relatively narrow. This approach, although developed for sand-bed channels, is now also the starting point for studying entrainment across a much wider range of bed grain sizes.

Dimensionless critical shear stress is commonly known as the *Shields number*. This empirically derived number typically varies between 0.03 and 0.08 for channels with $S < 0.03$, but can include values between 0.01 and 0.2 (Buffington and Montgomery, 1997). The wide variation in Shields number results from at least three factors (Ferguson, 2013; Yager and Schott, 2013). (1) Diverse studies use different criteria for defining the initiation of grain motion and different methods of measuring the initiation of motion. (2) Differences in the grain-size distribution and flow roughness in diverse channels produce a range of values for the Shields number. (3) The use of reach-averaged conditions results in a greater range of values.

It is also important to distinguish between the apparent Shields number and the actual Shields number. The *apparent Shields number* is calculated using Equation 4.2. The apparent Shields number is widely used because it is derived from relatively simple measurements of the reach-averaged shear stress and grain-size distribution. The apparent Shields number, however, is relatively simplistic because entrainment is typically not spatially or temporally uniform and thus is not accurately represented by a single shear stress for a particular grain, or a single grain size for a streambed (Yager and Schott, 2013). The actual Shields number of a particular grain reflects the force balance acting on the grain, which depends on the local particle arrangement and flow turbulence. Any given grain size thus has a range of possible critical shear stress values (Figure 4.1).

Research on entrainment in beds of mixed grain sizes focuses on the force balances acting on grains, the grain properties, and the turbulence parameters that influence motion. Biotic processes that influence entrainment have also recently received increasing attention.

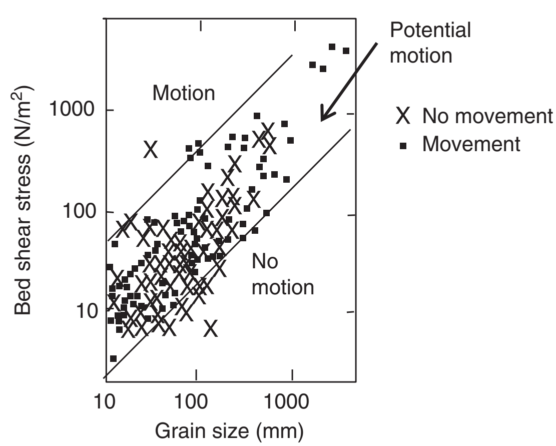

Figure 4.1 Bed shear stress versus grain size, showing range of entrainment (from Williams, 1983, Figure 1).

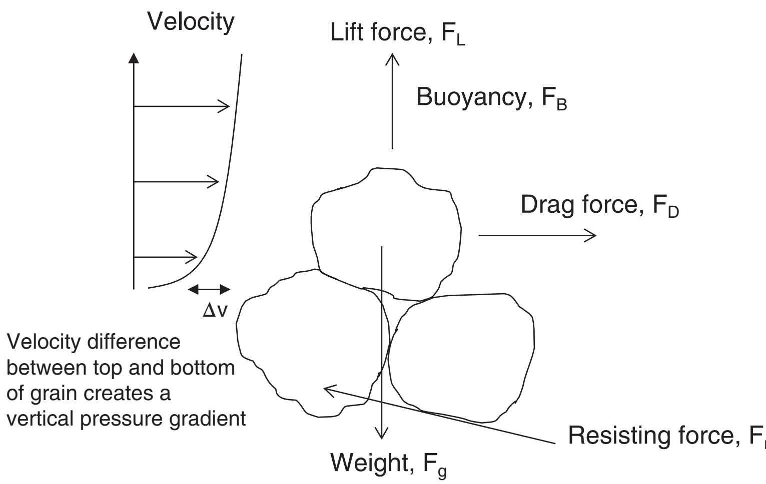

Figure 4.2 Schematic illustration of the forces acting on a non-cohesive grain under steady uniform flow on a nearly horizontal surface.

Forces acting on a grain

The balance of forces acting on a grain is schematically illustrated in Figure 4.2. The downstream and upward driving forces acting on the grain are the downstream component of particle weight F_g, buoyancy F_B, lift F_L, and drag F_D. The frictional resisting forces acting on the grain are F_r, the bed-perpendicular component of particle weight, and the friction angle. Entrainment occurs when the driving forces exceed the resisting forces. Each of these forces reflects the local turbulence and particle characteristics, including diameter, drag coefficient, friction angle (resisting pocket angle of a grain), and protrusion of the grain into the flow (Yager and Schott, 2013).

For spatially and temporally averaged flow conditions, the forces acting on a grain can be derived as follows (Wiberg and Smith, 1987; Yager and Schott, 2013). Drag force results from flow of water past a grain in the downstream direction and is

$$F_D = \frac{1}{2}\rho A C_D \left\langle v^2(z) \right\rangle = \frac{1}{2} A C_D \tau_b \left\langle f^2\left(\frac{z}{z_0}\right)\right\rangle \tag{4.3}$$

where C_D is the drag coefficient and is a function of the particle Reynolds number (the particle Reynolds number describes the ratio of kinetic to viscous forces applied on a moving particle), shape, size, orientation, and relative submergence; ρ is the water density; $\langle v^2(z) \rangle$ is the square of the fluid velocity averaged over the grain; τ_b is the reach-averaged boundary shear stress; $f(z/z_o)$ is a function that determines the form of the velocity profile around the grain; and A is the cross-sectional grain area, which varies with relative submergence and protrusion.

Lift force results from differences in velocity and pressure at the top and bottom of a grain and is

$$\begin{aligned} F_L &= \frac{1}{2}\rho A C_L \left(v_T^2 - v_B^2\right)^2 \\ &= \frac{1}{2} A C_L \tau_b \left\langle f^2\left(\frac{z_T}{z_0}\right) - f^2\left(\frac{z_B}{z_0}\right)\right\rangle \end{aligned} \tag{4.4}$$

where z_T and z_B are the heights of the top and bottom of the grain, respectively; C_L is the lift coefficient; and v_T and v_B are the flow velocities of the top and bottom of the grain, respectively. The larger the velocity and pressure gradients between the top and base of grains, the more important the lift force becomes.

The force exerted by particle weight is

$$F_g = \left(\rho_s - \rho\right) gV \tag{4.5}$$

where g is the gravitational acceleration and V is the particle volume. This force is the difference between the grain weight $(\rho_s gV)$ and buoyancy $(\rho_w gV)$, which is the amount of fluid weight displaced by the grain of interest. The net gravitational force is divided between the down slope (driving force) and bed-perpendicular (resisting force) components.

The driving forces can then be written as

$$F_D + F_L \tan\phi + F_g \sin\beta \tag{4.6}$$

where ϕ is the friction angle and β is the channel slope. The friction angle is the angle through which a grain will pivot when it is dislodged in the downstream direction.

The resisting force is given by

$$F_g \cos\beta \tan\phi \tag{4.7}$$

For entrainment, Equations 4.6 and 4.7 can be equated, Equations 4.3 and 4.4 can be substituted for F_D and F_L, respectively, and the equation can be solved for the critical shear stress to yield

$$\frac{\tau_c}{(\rho_s - \rho)gD} = \frac{2\mathrm{V}\,(\tan\phi\cos\beta - \sin\beta)}{C_D A D \left\langle f^2\left(\frac{z}{z_0}\right)\right\rangle \left(1 + \left[\frac{F_L}{F_D}\right]\tan\phi\right)} \tag{4.8}$$

This equation requires an assumed velocity profile. The logarithmic profile is commonly used. Other representations of the force balance use different forces and alter the parameterization of each force (Yager and Schott, 2013). Bagnold (1977), for example, proposed using stream power per unit area rather than shear stress to evaluate entrainment and transport.

Although Equation 4.8 addresses the basic forces acting on a grain, it is in practice cumbersome to apply to real channels with a range of grain sizes. This reflects the substantial scatter in actual, site-specific values of individual forces associated with grain properties and flow turbulence.

Grain properties

The grain properties that influence entrainment include grain size, shape, orientation, packing, and imbrication. As noted previously, the original Shields curve of dimensionless shear stress and sediment transport rate was empirically derived for relatively uniformly sized sediment. Larger grains have a higher critical shear stress because of their greater weight, and weight-driven transport represents one end-member for entrainment from a bed of mixed grain sizes. Equal mobility transport represents the other end-member.

Equal mobility can be defined as the same critical shear stress (τ_c) for all grain sizes (Parker et al., 1982). Under this condition, τ_c^* decreases with increasing grain size because coarser particles protrude into the flow and are easier to move than on a bed of uniform grain size, whereas finer particles are hidden and therefore harder to move than on a bed of uniform grain size. Equal mobility can also be defined as occurring when the grain-size distribution of the bed load equals that of the channel bed (Parker and Toro-Escobar, 2002).

Under conditions of intermediate mobility between the two end-members of weight-driven and equal mobility, fine sediment is likely to be more mobile than coarse sediment, but less mobile than it would be on a bed of uniform fine particles. This is known as *selective entrainment* (Parker, 2008).

The competing effects of grain weight and hiding can be represented by hiding functions that estimate the influence of mixed sediment sizes and flow conditions on entrainment, as in this example from Parker et al. (1982)

$$\frac{\tau_{\mathrm{ci}}^*}{\tau_{\mathrm{c50}}^*} = \left(\frac{D_{\mathrm{i}}}{D_{50}}\right)^b \tag{4.9}$$

where τ_{ci}^* and τ_{c50}^* are the dimensionless critical shear stresses for the ith grain size (D_{i}) and the median grain size (D_{50}), respectively, and b indicates the relative importance of grain weight and hiding effects. When grain weight dominates, $b = 0$. When hiding effects dominate and create equal mobility, $b = -1$ (Yager and Schott, 2013).

Because hiding functions are derived from site-specific empirical measurements, they are not transferable between locations (McEwan et al., 2004). Accounting for hiding effects when trying to quantitatively predict entrainment becomes progressively more important, however, as the grain-size distribution grows wider and entrainment of the smallest grains is more influenced by hiding.

In addition to grain size, grain protrusion, p, grain exposure, e, and friction angle, ϕ, influence entrainment. *Protrusion* is the distance a particle protrudes above the mean bed elevation. *Exposure* is the distance a grain extends above or below the neighboring grains. *Friction angle*, also known as pivoting angle, would be about 30—32 degrees for uniform sediments (Robert, 2003). In heterogeneous sediments, the angle through which a grain must pivot to be entrained varies significantly, depending on the size distribution of the surrounding grains. Protrusion, exposure, and friction angle can vary with relative grain size, shape, and packing, and are typically empirically estimated.

Beds with increasingly wide distributions of grain size have increasingly larger effects from protrusion, exposure, and hiding. Large, protruding, immobile grains can be particularly important in creating higher values of dimensionless critical shear stress in steep, shallow flows. The additional resistance created by these grains and by immobile bedforms reduces the shear stress available for grain entrainment, so that the Shields number increases with slope (Ferguson, 2013).

Turbulence

Near-bed turbulent bursts, sweeps, and inward and outward interactions can influence entrainment by inducing fluctuations in the pressure and velocity fields around a particle and thus influencing drag and lift forces (Nelson et al., 1995). The relationship between local turbulent forces and grain entrainment is quite complex and sediment motion can be dominated by drag, lift, or some combination of both forces (Schmeeckle et al., 2007). Both magnitude and duration of turbulent fluctuations influence entrainment (Diplas et al., 2008).

Spatial and temporal variability in bed texture and hydraulic forces make it effectively impossible to predict exactly when a particular grain will be entrained. Entrainment of a particular grain or a population of mixed grain sizes is instead commonly described in terms of a range of critical values for velocity, shear stress, or some other parameter.

Biotic processes

Biotic processes can also influence entrainment (Riggsbee et al., 2013). Instream wood creates large-scale hiding effects and alters the distribution of pressure and velocity near the bed. Microbial mats and macroinvertebrates such as net-spinning caddisfly larvae can increase the cohesion of sand-size and finer sediments, and thus increase the shear stress necessary for entrainment (Statzner et al., 1996). Algae and microorganisms can facilitate deposition of calcium carbonate, which limits sediment mobility and enhances the formation of quasi-permanent steps and pools (Marks et al., 2006). Bioturbation by invertebrates and the building of spawning mounds or depressions by fish can alter surface grain-size distribution and topography (Statzner and Sagnes, 2008).

The magnitude of these effects varies in part with the abundance of the biota. Instream wood was historically widespread and abundant along rivers in forested regions and undoubtedly strongly influenced sediment dynamics, but centuries of removing wood from channels has diminished these effects. Bed sediment dynamics are substantially influenced by spawning activities in rivers that still have abundant salmon populations (Hassan et al., 2008), although such rivers are now rare.

4.1.3 Erosion of cohesive beds

Channels formed in cohesive material can have a thin, continuous or discontinuous veneer of unconsolidated sediment along the bed. The cohesive material underlying this alluvium limits the rate and manner of bed and bank erosion and exerts a strong influence on cross-sectional geometry and hydraulics during large discharges because the boundary is not readily erodible. Cohesive material in this context includes channel boundaries with sufficient silt and clay to create strong interparticle cohesion and limit the detachment of individual silt or clay particles, and bedrock. Cohesive boundaries are not necessarily more resistant to erosion than alluvial boundaries. A very soft siltstone or sandstone may have a lower erosional threshold than an alluvial channel formed in boulders. In general, however, cohesive material has a higher erosional threshold than non-cohesive alluvium.

Because individual particles are not readily detached from cohesive material, sediment entrainment is not the only process by which channels in cohesive beds erode, although sediment entrainment can be an important component of other erosive processes. The primary erosive processes in indurated, cohesive material (bedrock) are corrosion, cavitation, abrasion, and quarrying (Whipple et al., 2013).

Corrosion is the chemical dissolution of bedrock. Most chemical weathering occurs outside the channel, where water moves more slowly through or past the rock matrix. Chemical weathering can be effective in eroding carbonate rocks along channels, however, and in weakening other rock types, particularly those with carbonate cement (Springer et al., 2003). Chemical weathering can thus make the bedrock more susceptible to processes of physical erosion (Hancock et al., 2011).

Cavitation involves shock waves generated by the collapse of vapor bubbles in a flow with rapidly fluctuating velocity and pressure. The minute irregularities along channel boundaries caused by joints or crystal or grain boundaries in the bedrock can facilitate the formation and collapse of vapor bubbles. Although this effect might sound insignificant, millions of bubbles imploding and sending out shock waves during a flood can very effectively weaken and erode bedrock, a phenomenon common on the concrete spillways of dams (Barnes, 1956; Eckley and Hinchliff, 1986; Wohl, 1998). The importance of cavitation relative to the processes of quarrying and abrasion remains largely unquantified, however, for natural channels.

Quarrying refers to the entrainment of blocks at least partially detached from the surrounding bedrock, typically along joints. Lift force is particularly important in quarrying, and the large lift forces generated in shallow, swift flow can effectively quarry blocks much greater than 1 m across (Wohl, 1998; Whipple et al., 2000; Chatanantavet and Parker, 2009; Dubinski and Wohl, 2012). Quarrying can be the dominant mechanism of bedrock channel erosion in strongly jointed bedrock.

Abrasion is abrasive erosion of bedrock by clasts carried in the flow, and *macroabrasion* is sometimes used to refer to the combined effects of particle impacts that fracture bedrock into fragments that can be quarried (Chatanantavet and Parker, 2009). The effectiveness of abrasion depends on (i) the relative hardness of the clasts in transport and the cohesive boundaries, (ii) the amount of sediment in transport, and (iii) the frequency and duration of flows during which clasts able to create abrasion are in transport. Abrasion has a nonlinear relation with the amount of sediment available. Very low sediment supply limits the tools for abrasion, but high sediment supply can cover and protect the bed from abrasion (Sklar and Dietrich, 2004) (Figure S4.4).

Bed erosion in cohesive materials typically does not occur evenly across a cross section or throughout a reach, and bedrock channels seldom have a planar bed. Instead, erosion tends to occur most actively over a fraction of the bed width that varies with bed load supply and transport capacity (Finnegan et al., 2007). This results in features such as inner channels and variable bed gradients (Baker, 1978; Wohl, 1998).

Just as alluvial bedforms such as ripples and dunes both reflect and influence boundary roughness, velocity, and turbulence, so feedbacks among boundary roughness, rate of erosion, velocity and turbulence, and sediment mobility influence the location and style of erosion in bedrock channels (Goode and Wohl, 2010b). A distinctive result of localized erosion is the creation of *sculpted forms* such as potholes, grooves, and undulating walls (Richardson and Carling, 2005). The location and dimensions of some sculpted forms can be used to infer hydraulics during formative flows (Wohl, 2010).

Equations for processes of bedrock erosion, and further information on processes of bedrock erosion, are in Supplemental Section 4.1.3.

Erosive processes in cohesive but non-indurated materials, such as silt and clay, are similar to those in bedrock channels, but are also influenced by upward-directed seepage and matric suction. *Upward-directed seepage* is seepage directed vertically upward. This leads to static liquefaction, which creates an additional driving force. *Matric suction* occurs when negative pore water pressure above the water table increases the apparent cohesion of a soil; this is also called matrix suction. Matric suction creates an additional force of resistance to bed erosion (Simon and Collison, 2001).

Much of the research on erosion of cohesive, non-indurated materials focuses on stream banks rather than the streambed, and is discussed in Section 4.3.

4.2 Sediment transport

4.2.1 Dissolved load

The sediment carried in solution within a river is sometimes ignored as a component of sediment transport, but can be substantial in some rivers. A survey of the world's large rivers indicates anywhere from 2% of the total load carried in solution in the Huang He River of China to 93% in Canada's St. Lawrence River (Knighton, 1998).

Solute concentration is highest in water entering a channel through subsurface pathways in which generally slower rates of flow provide longer reaction times with the surrounding matrix. Solute concentration typically declines with increasing discharge as water moving more rapidly and along surface flow paths enters a river. Total dissolved load continues to increase with discharge, but much of the dissolved load is carried by relatively frequent flows.

Solute concentration also displays *hysteresis effects*, with higher solute concentrations on the rising limb than on the falling limb as a result of mobilization of soluble material that accumulates prior to the flood (Walling and Webb, 1986). Rivers with strongly seasonal flow regimes typically have an annual flush of high dissolved loads, as in the rising limb of a snowmelt-dominated river.

A global survey of 370 rivers indicates that solute concentration, C, varies with discharge, Q, as

$$C = aQ^b \tag{4.10}$$

where b is mostly 0 to −0.4, with a mean value of −0.17 (Walling and Webb, 1986).

As noted in Chapter 2, the primary constituents of dissolved load are the dissolved ions HCO_3^-, Ca^{2+}, SO_4^{2-}, H_4SiO_4, Cl^-, Na^+, Mg^{2+}, and K^+; dissolved nutrients N and P; dissolved organic matter; dissolved gases N_2, CO_2, and O_2; and trace metals (Berner and Berner, 1987; Allan, 1995). The sum of the concentrations of the dissolved major ions is known as the *total dissolved solids* (TDS). An average natural value for rivers is 100 mg/L, and pollution adds on average another 10 mg/L. Ca^{2+} and HCO_3^- from limestone weathering tend to dominate TDS (Berner and Berner, 1987). Because dissolved fluxes in rivers partly reflect rates of continental weathering, which in turn reflect rates of tectonic uplift, the largest contemporary fluxes occur in rivers draining the Himalayan and Andean mountains and the Tibetan Plateau (Raymo et al., 1988).

Reach-scale dissolved load can also strongly reflect interactions between surface and hyporheic water. Downwelling into the hyporheic zone transfers solutes and surface water rich in dissolved oxygen, whereas upwelling flow can transfer nutrients to streams (Tonina and Buffington, 2009). Hyporheic exchange exposes solutes in stream water, including nutrients, to alternating anoxic and oxic zones in the bed that are composed of geochemically reactive sediments and microbial communities (Lautz and Siegel, 2007). Anoxic zones that indicate the presence of sulfate, iron, and manganese reduction typically occur upstream of streambed structures, as in low-velocity pools. Oxic zones that reflect the production of nitrate are typically downstream of bed structures, as in turbulent riffles (Lautz and Fanelli, 2008). These zones can enhance biogeochemical reactions and increase nutrient utilization, and may be critical to maintaining stream water quality (Hancock et al., 2005).

Patrick (1995) reviewed the importance of river chemistry to aquatic organisms. Dissolved organic matter provides an important energy and nutrient source. Ca, Mg, oxidized S, N, and phosphates, along with small amounts of Si, Mg, and Fe, are desirable for many species. The pH of the water affects the solubility and availability of elements.

Nitrogen and carbon are important nutrients in river water. Rivers are major conduits for nitrogen and carbon transport, but rivers can also remove and transform dissolved nitrogen and carbon in transport (Hall and Tank, 2003). The details of nitrogen and carbon transport versus removal provide a nice example of how physical, chemical, and biochemical processes interact in rivers to govern transport of solutes.

Of the total nitrogen loss from the land, only about 18%–20% is carried to the oceans by rivers (Van Breemen et al., 2002) because of removal and transformation en route. Rates of dissolved nitrogen uptake from river water are especially high in shallow streams with algae and microbes in attached biofilms (Hall and Tank, 2003), but riverine processing of nitrogen is heterogeneous in time and space. Examples of biogeochemical hot spots in which accelerated chemical reactions occur include anoxic zones beneath riparian environments (Lowrance et al., 1984) or the convergence of ground and surface waters in hyporheic zones (Harvey and Fuller, 1998). Examples of biogeochemical hot moments include rare high flows that occupy secondary channels in arid environments (McGinness et al., 2002) or snowmelt that enhances leaching of dissolved nutrients in high-elevation catchments (Boyer et al., 2000).

Much of the nitrogen removal within rivers occurs in stream sediments (Nihlgard et al., 1994) and riparian zones (McClain et al., 1999) through the activities of benthic (bottom-dwelling) organisms such as bacteria, algae, and insects, and riparian soil organisms and plants that take up nitrogen. Any source of fluvial complexity—logjams, beaver dams, channel margin irregularities, hyporheic zones, and shallow overbank flow across floodplains—that promotes flow separation and retention of organic matter can substantially increase nitrogen uptake. By retaining organic matter, these sites provide biota an opportunity to access and ingest the organic matter (Naiman et al., 1986; Fanelli and Lautz, 2008). Partly for this reason, river management increasingly emphasizes protection or restoration of physical channel complexity, as well as protection or restoration of riparian corridors that can "buffer" channels from excess nitrogen.

Human activities, including use of inorganic fertilizer and emissions from fossil-fuel combustion, have dramatically increased nitrogen inputs to watersheds (Boyer et al., 2006). Simultaneously, river engineering that limits channel boundary complexity and retention, as well as channel–floodplain connectivity, has reduced the ability of many rivers to process nitrogen inputs. The result is excess nitrogen loads and eutrophication in coastal areas (Boyer et al., 2006). Many large river systems are reaching their limits for processing excess nitrogen. Preindustrial nitrogen fluxes were greatest from the largest rivers, such as the Amazon. Post-industrial fluxes are greatest from rivers in the industrialized zones of North America, Europe, and southern Asia (Green et al., 2004).

Organic matter can be present in river water in particulate and dissolved forms (Berner and Berner, 1987). Particulate organic matter enters rivers as fossil carbon from sedimentary rocks and as carbon from contemporary soils and vegetation. Carbon inputs can be gradual and continuous. Carbon inputs can also be episodic, as in active mountain belts where sediment yield to rivers is dominated by landslides triggered by tropical cyclones, which also generate floods that result in large riverine fluxes of carbon to the ocean (Hilton et al., 2008a, 2008b, 2011a, 2011b).

Dissolved organic matter includes different classes of organic compounds that differ in reactivity and ecological role, as well as varying in quantity with time and space in response to seasonal variation and precipitation inputs. Dissolved organic matter is usually expressed as dissolved organic carbon (DOC). Values in river waters are typically

2–15 mg/L but can reach 60 mg/L in rivers draining wetlands (Drever, 1988). DOC varies with the size of the river, the climate, and vegetation (Thurman, 1985). DOC values are typically high for temperate and tropical rainforests and taiga, for example, and low for arid and semiarid environments.

DOC is highly reactive and influences riverine ecosystems by controlling microbial food webs (Kaiser et al., 2004). The efficiency with which rivers retain and oxidize organic carbon depends on the presence of microbial communities in response to riverine features that increase the residence time of organic molecules in transport, as explained above for nitrogen (Battin et al., 2008, 2009). Simply put, storage of organic matter supports microbial communities, which retain and oxidize organic carbon.

Substantial differences in carbon dynamics are present between rivers with well-developed floodplains, which exchange more than half of the recent biogenic carbon in transport with floodplain carbon, and small mountainous rivers that transport carbon directly to the ocean (Galy et al., 2008). Even small mountainous rivers can include segments with substantial carbon storage, however, if physical channel complexity creates flow obstructions and enhanced channel–floodplain connectivity (Wohl et al., 2012c). Consequently, river management that facilitates maintenance or restoration of channel complexity has numerous benefits, including carbon and nitrogen retention and increased diversity of aquatic habitat.

DOC is an important component of the global carbon cycle. An estimated 2.7 petagrams (Pg) of carbon are eroded from terrestrial sources and delivered to freshwater environments each year (Aufdenkampe et al., 2011). Of this quantity, 0.6 Pg is buried in sedimentary reservoirs such as floodplains or deltas. Another 1.2 Pg is released to the atmosphere as a result of metabolism and respiration of riverine organisms. The remaining 0.9 Pg is carried by rivers to the ocean. River dynamics strongly influence the partitioning between sedimentary reservoirs, release to the atmosphere, and transport to the ocean.

Trace metals typically occur at concentrations less than 1 mg/L in river waters. Trace metals can be derived from rock weathering, or from human activities that include mining, burning fuels, smelting ores, or disposing of waste products (Drever, 1988). Sediments contaminated with trace metals associated with human activities are commonly referred to as *legacy sediments* (as are other types of deposits that reflect past human activities). These contaminated sediments create a long-term toxic legacy. Trace metals receive most attention as contaminants occurring at concentrations above background levels or hazards occurring at levels potentially harmful to organisms. Examples of trace metals that commonly create pollution are arsenic, cadmium, chromium, copper, lead, mercury, nickel, selenium, and zinc, each of which is very toxic and biochemically accessible to living organisms. Because many trace metals are adsorbed to fine sediment, the mobility and storage of the fine sediment strongly influence the spread of trace metals through a river network and the bioavailability of the metals. Metals adsorbed to silt and clay traveling in suspension or resting on the surface of the streambed, for example, are more readily ingested by aquatic organisms than are metals adsorbed to fine sediment deeply buried on a floodplain.

As with other aspects of hydrology, cold-region rivers and rivers of the humid tropics have some distinctive chemical characteristics. Cold-region rivers in alpine or permafrost terrains with high levels of impermeable surface area may be more influenced by precipitation chemistry than other types of rivers with greater infiltration and deeper, slower downslope pathways of water (Meixner et al., 2000). Selective weathering in glaciated catchments can allow more reactive minerals to contribute disproportionately to dissolved load relative to their abundance in the local rock. Chemical denudation rates can be higher for glaciated areas than for adjacent non-glaciated catchments. Both seasonal snowmelt and glacial melt can create large temporal and spatial variations in stream water chemistry as the source and contact time of melt water with sediment and rock change through the melt season (Taylor et al., 2001; Anderson et al., 2003). Total organic carbon export from cold-region rivers is likely to increase under global warming because of accelerated decomposition of soil organic matter as the seasonal duration of frozen soil decreases (Trumbore and Czimczik, 2008).

Rivers of the humid tropics have warmer water, higher annual exports of dissolved constituents (Lyons et al., 2002), and lower seasonal variability in water temperature and chemistry than rivers at higher latitudes (Scatena and Gupta, 2013). Carbon exports occur primarily as DOC, and both DOC and particulate carbon exports can increase significantly when tropical storms create widespread defoliation and/or landsliding (Hilton et al., 2008a, 2008b). Consequently, humid tropical rivers can create important global sinks of carbon by transporting carbon to oceanic burial (Goldsmith et al., 2008).

4.2.2 Suspended load

The particulate sediment transported by rivers can be subdivided in different ways. One distinction is between wash load and bed-material load, another is between suspended load and bed load. The distinction between wash and bed-material load reflects different mechanics of transport, but the coarsest grain size moving as wash load is not easy to measure directly, blurring this distinction. The distinction between suspended and bed load primarily reflects measurement technology, with some forms of bed load transport (saltation) being transitional between the two end-members.

Wash load is the finest size fraction of the total sediment load, typically grains with intermediate diameter ≤0.062 mm. Wash load consists of particles typically not found in large quantities on the bed surface. These particles have settling velocities so small that they move at approximately the same velocity as the flow and only settle from suspension when velocity declines substantially. Wash load is vertically mixed by turbulent flows, so that concentration varies little with flow depth. Because relatively little energy is needed to transport wash load, transport rates typically reflect sediment supply rather than flow energy. Much of the wash load comes from bank erosion and surface erosion across the drainage basin.

Interparticle forces are greater than the gravity force among portions of the wash load finer than 0.004 mm in diameter, and these very fine sediments are cohesive. Cohesive sediments typically travel as aggregated or flocculated material, rather than as single grains (Kuhnle, 2013). Unlike coarser sediments, for which transport may be limited by stream energy (and which are thus known as *transport limited*), fine cohesive sediments are more likely to be *supply limited*, such that the amount in transport is limited by the amount supplied to the flow. Under supply-limited conditions, the actual *transport rate* is likely to be less than the *transport capacity*, which reflects what the flow is capable of transporting.

Bed-material load includes grains typically coarser than 0.062 mm. These grains move either in contact with the bed by rolling, sliding, or saltating as *bed load*, or in suspension just above the bed, with concentration declining upward from the bed. Bed-material load can also be subdivided into suspended load and bed load. If sediment is divided into only suspended load and bed load, then the suspended load includes wash load.

Non-cohesive sediment will remain in suspension if the strongest vertical velocity fluctuations exceed the particle fall velocity. Close to the bed, the root mean square of the vertical velocity fluctuations reaches a maximum that is approximately equal to the magnitude of the *shear velocity*, v_*, which provides an approximate criterion for suspension (Kuhnle, 2013)

$$v_* = \sqrt{\frac{\tau_0}{\rho}} \tag{4.11}$$

where τ_0 is the bed shear stress and ρ is water density. Non-cohesive suspended sediment concentration typically decreases with distance away from the bed in a manner described using the Rouse equation

$$\frac{C}{C_a} = \left(\frac{(a)\,(d-\gamma)}{(\gamma)\,(d-a)}\right)^Z \tag{4.12}$$

where C is the concentration of sediment at a distance γ above the mean bed elevation, C_a is the concentration of sediment at the reference level a above the bed, d is flow depth, and the exponent, Z, known as the *Rouse number*, is defined as

$$Z = \frac{\omega_s}{\kappa u_*} \tag{4.13}$$

where ω_s is the particle fall velocity, κ is the von Karman constant, and u_* is the shear velocity (Kuhnle, 2013).

The Rouse number is basically the ratio of the grain fall velocity to the strength of the flow. The relation between the Rouse number and the distribution of suspended sediment through the water column (Figure 4.3) reflects the fact that, as the value of the flow strength (u_*) increases relative to the fall velocity of the sediment grains, the resulting gradient of sediment concentration through the flow depth grows less steep (Kuhnle, 2013).

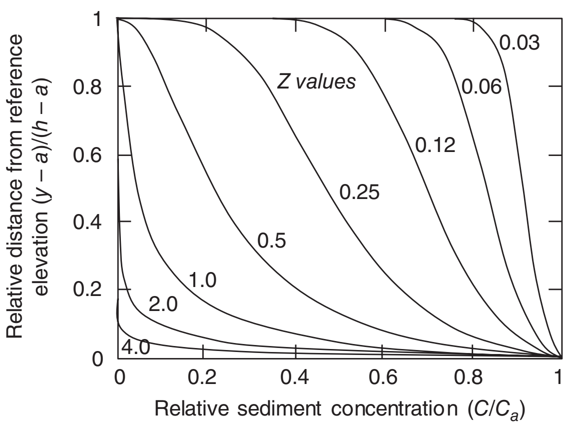

Figure 4.3 Diagram illustrating the effect of a range of Rouse numbers (Z values) on the prediction of suspended sediment concentration profiles. The variables on the x and y axes are as follows: a is a reference level above the bed, h is depth of flow, C is the concentration of sediment at a distance y above the mean bed elevation, and C_a is the concentration of sediment at the reference level a. Z values reflect the ratio of grain fall velocity to the strength of the flow. Greater values of flow strength relative to the fall velocity (smaller Z values) of the grains result in a less steep gradient of sediment concentration with flow depth—that is, greater concentrations of suspended sediment high in the flow than at large Z values (from Kuhnle, 2013, Figure 2).

Large suspended sediment concentrations commonly occur during high discharge. Using an equation similar to Equation 4.10, the exponent b is generally in the range of 1 to 2 (Walling and Webb, 1986). Wolman and Miller (1960) first suggested that most suspended sediment is carried by relatively frequent flows with a recurrence interval of 1–2 years, although the magnitude of flow carrying the greatest proportion of annual suspended sediment load—also known as the *effective discharge*—varies widely among catchments.

Effective discharge is usually a relatively frequent flow in larger basins. Working in Canada's Saskatchewan River basin, for example, Ashmore and Day (1988) found that small headwaters basins tended to have effective discharge during the most extreme events with annual flow duration much less than 1%, whereas larger rivers had most suspended sediment transported during frequent events with durations often greater than 10%.

Supplemental Section 4.2.2 includes more information about suspended sediment concentrations in relation to flow magnitude and recurrence interval.

Most rivers display some form of hysteresis for suspended sediment, but the details can vary widely. Williams (1989) identified three forms of hysteresis (Figure 4.4). One form is a clockwise loop in which the peak sediment concentration precedes the peak discharge because available sediment is depleted before runoff peaks, a situation more common in small basins. A second form of hysteresis is an anticlockwise loop in which peak sediment concentration lags the peak discharge, a situation more common in large basins. The third form of hysteresis is a figure-of-eight loop in which peak sediment concentration precedes peak discharge, but the shape of the sediment output is skewed relative to the flood peak. This can occur where suspended sediment increases more rapidly than discharge, and thus peaks first, but post-peak sediment availability and transport are sufficiently high to create sediment concentrations that decrease more slowly than water discharge.

Sources of suspended sediment are typically not uniformly distributed across a drainage basin, particularly in larger basins. This is dramatically

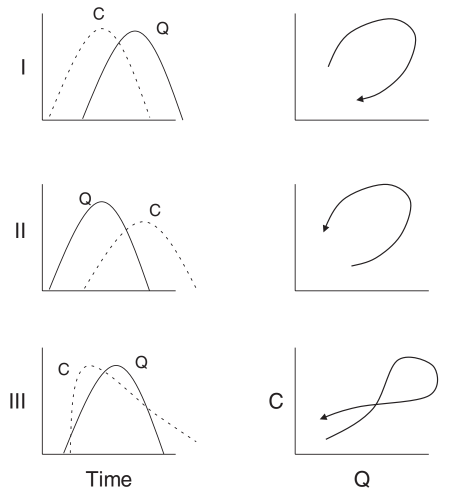

Figure 4.4 Three types of hysteresis in suspended sediment—discharge relations when *C* (concentration) and *Q* (discharge) are not synchronous (from Williams, 1989, Figures 3, 5, and 7).

illustrated by Bob Meade's "sedigraphs" for drainage basins such as the Mississippi River in North America (Meade, 1996) (Figure 4.5). The majority of discharge in this drainage comes from the humid temperate Ohio River basin and the headwaters of the Mississippi, whereas the majority of suspended sediment comes from the semiarid Missouri River basin. The relative importance of different sub-basins as sources of suspended sediment reflects sediment yields as influenced by factors such as rock type, topography, climate, and land use. Sediment budgets for the Amazon River basin show similar spatial disparities as those for the Mississippi, with tributary basins draining highlands contributing substantial suspended sediment, whereas other tributaries contribute negligible amounts.

Suspended sediment dynamics along the Amazon also indicate the importance of the floodplain and channel banks. Much of the suspended sediment in transport during the rising and peak stages of the annual flood is deposited in the floodplain, where the sediment remains until returned to fluvial transport via bank erosion (Dunne et al., 1998).

Suspended sediment can also be strongly influenced by land use that disrupts land cover and increases erosion from upland surfaces via processes such as rainsplash, sheetwash, and rilling. An example comes from the 539 km^2 Lago Loíza basin of Puerto Rico. Completion of a dam in 1953 created the lake (Lago Loíza). Gellis et al. (2006) used historic records of suspended sediment load on incoming rivers, repeat bathymetric surveys of the lake, sediment cores from the lake, precipitation records, and land use maps to evaluate potential controls on suspended sediment yield to the reservoir. They found that upland clearing for agriculture resulted in increased sheetwash, rilling, and downstream sediment yields, as did subsequent urbanization during the construction phase. Some of the sediment mobilized from uplands during agricultural development was stored along channels and floodplains for several decades. This sediment was remobilized when second-growth forest took over croplands, so that downstream sediment yields remained high despite upland reforestation.

Potential effects of land use on suspended sediment are covered in more detail in Supplemental Section 4.2.2.

Suspended load is the dominant component of sediment transport on many rivers, particularly rivers draining areas larger than about 5000 km^2 (Kuhnle, 2013). Suspended load is also the most commonly measured component of total sediment load. Globally, suspended sediment accounts for about 70% of total fluvial sediment transport to the oceans. Small, mountainous basins produce much greater yield per unit area (tons/km^2), but large drainages produce the greatest total sediment output.

Suspended sediment can influence hydraulics, total sediment transport, and channel form in diverse ways. Large concentrations of suspended sediment can increase flow viscosity, reduce settling velocities, and thus decrease turbulence and facilitate the transport of coarser grains than in clear water (Simons et al., 1963). Suspended sediment can also create turbidity that limits photosynthesis for aquatic plants. Infiltration of fine suspended sediment into an alluvial streambed can increase bed cohesion and reduce porosity and permeability of the sediment matrix. Reduced porosity and

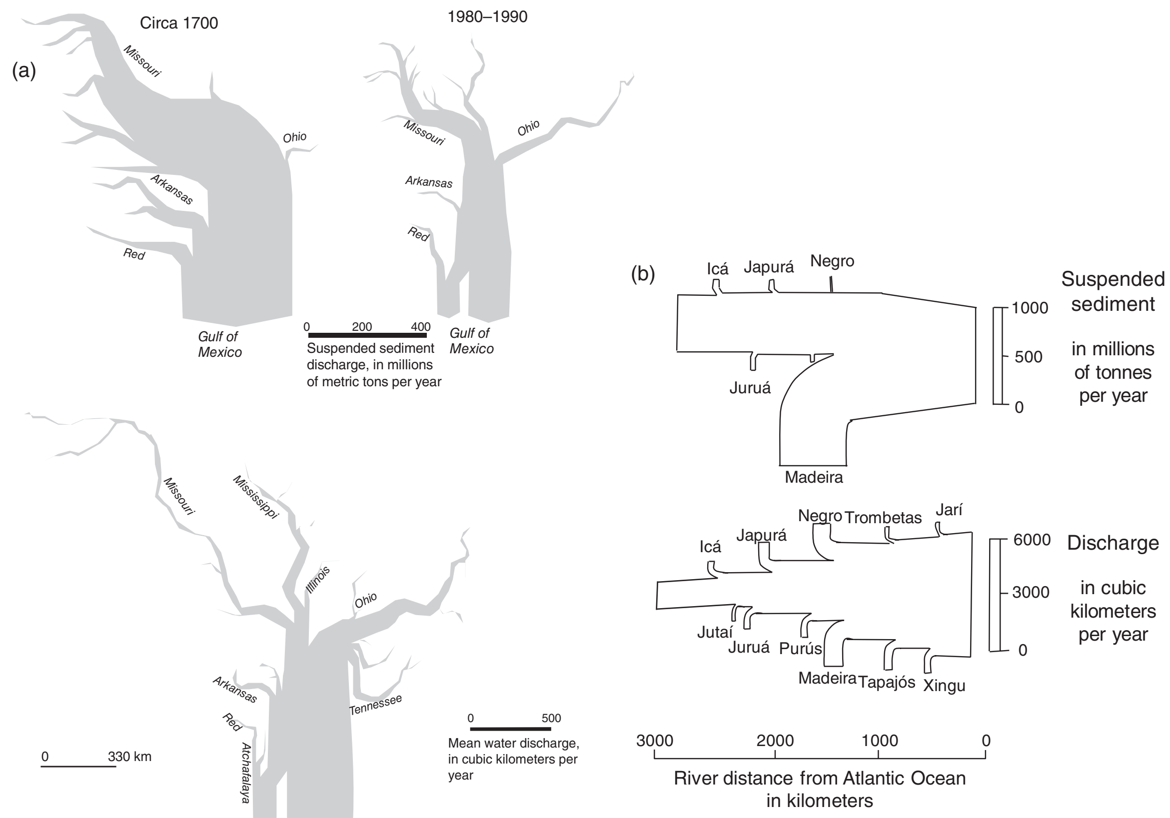

Figure 4.5 (a) Schematic illustration of historical and contemporary contributions to suspended sediment discharge (above) and water discharge (below) of the Mississippi River (from Meade, 1996, Figures 5 and 6). The majority of the suspended sediment carried in the Mississippi continues to come from the Missouri River, although this value has declined substantially through time as successively more large dams have been built along the Missouri, resulting in substantial sediment storage in reservoirs. The Ohio River is the single largest contributor of discharge to the Mississippi. (b) Schematic representation of downstream changes in suspended sediment and discharge in the Amazon River (from Meade, 2007, Figure 4.6, p. 53). The Rio Negro and Rio Madeira contribute similar magnitudes of discharge to the Amazon, but the Rio Madeira contributes substantially more suspended sediment.

permeability can in turn reduce hyporheic exchange and the ability of stream organisms from aquatic insects to fish embryos to survive within the bed matrix. High suspended sediment concentrations can correlate with high rates of overbank deposition and both lateral and vertical accretion of floodplains. High suspended loads can also correlate with stream banks with a greater percentage of silt and clay, greater bank cohesion, and relatively narrow, deep cross sections (Schumm, 1960).

A variety of pollutants, including heavy metals, pesticides, and excess nutrients, can travel adsorbed to cohesive fine sediments, so that the transport and deposition of fine sediments can strongly influence the distribution of contaminants within the river corridor. Nineteenth century silver and gold mining using a mercury-amalgamation procedure led to release of thousands of kilograms of mercury-contaminated tailings into the Carson River in Nevada, USA. Mercury, which is associated with fine-grained sediment fractions (<63 μm), has subsequently been dispersed downstream for tens of kilometers via overbank deposition during floods (Miller et al., 1999).

Supplemental Section 4.2.2 discusses measurement and modeling of suspended load.

4.2.3 Bed load

Although the contribution of bed load to total sediment transport is relatively small except in mountainous river networks, bed load dynamics strongly influence river process and form. The movement of bed load influences the stability of the bed and disturbance regime of aquatic habitat. Moving bed material influences flow resistance and river morphology, as well as lateral channel movement and floodplain formation. In tectonically active regions, the rates of channel incision and landscape evolution fundamentally reflect the rate at which bed load is moved out of the channel network.

Grain motion within bed load occurs via rolling and sliding at lower flows, saltation at intermediate flows, and sheetflow during high flows, although the three types of motion can coexist (Haschenburger, 2013). Grains rolling or sliding along the bed can make microleaps across distances smaller than the grain diameter, or begin to saltate. *Saltation*, from the Latin for "leap," describes grains launched into the water column that follow a predictable, ballistic trajectory before once more contacting the bed. The maximum saltation height, which can be up to ten times the grain diameter, defines the upper extent of the bed load layer. *Sheetflow* occurs when grains move in multiple granular sheets with a grain concentration approaching that of the underlying stationary bed. (Sheetflow is also used to refer to sediment movement under very shallow flows on hillslopes.)

The ratio of dimensionless shear stress to critical dimensionless shear stress (τ_*/τ_{*c}) correlates with type of sediment transport. Bed load typically dominates where τ_*/τ_{*c} is ~1–3, bed load and suspended load occur where τ_*/τ_{*c} is between 3 and 33, and suspended load dominates where τ_*/τ_{*c} exceeds ~33 (Church, 2005).

Most bed load comes from the streambed and lower banks. Bed load flux depends on grain velocity and concentration. At low concentrations, bed load particles collide primarily with stationary bed grains. Collisions between mobile grains become more frequent as bed load concentrations increase. Bed load grains are supported by grain collisions in a granular-fluid flow at high concentrations. Consequently, the mechanics of bed load motion vary because fluid–grain and grain–grain interactions change with concentration (Haschenburger, 2013).

As in the discussion of bed sediment characterization, it is useful to distinguish bed load in predominantly sand-bed channels from bed load in channels with coarser substrate and a wider range of grain sizes. For both types of channels, bed load movement is typically episodic and spatially heterogeneous across the channel and downstream.

In sand-bed channels, much of the bed load moves as migrating bedforms such as ripples, dunes, and antidunes, which are treated in the next section. Transport rate reflects the energy of the flow, because the individual sand grains are relatively mobile and transport is not typically limited by sediment supply. Sand-bed channels commonly exhibit bed scour during the rising stages of flow, but then fill to approximately pre-flood levels on the falling stage (Leopold et al., 1964).

In channels with substrate coarser than sand, bed load movement is more episodic in time and space, and is typically limited by sediment supply. Supply limitations on transport, as well as large spatial and temporal variations in local flow field and boundary configuration, make it difficult to predict bed load transport using foundational equations of bulk sediment transport developed for uniformly sized grains (typically sand sizes) in flumes. Consequently, research has shifted toward using two-phase equations that differentiate sand and gravel. Gravel-bed channels can also exhibit scour during the rising limb of a flood and fill during the falling limb.

Bed load in channels with coarse-grained substrate: coarse surface layer

In streams with coarser bed material, flow energy is important, but the finer portion of the grain-size distribution can also be supply limited by the existence of a coarse surface layer. Many channels with relatively coarse, poorly sorted grain-size distributions have at least slightly finer-grained sediments immediately below the coarse surface layer (Figure 4.6). The coarser grain sizes at the surface increase the

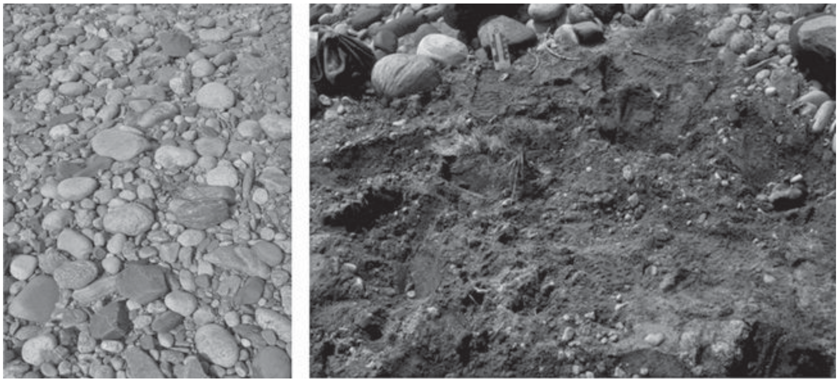

Figure 4.6 Coarse surface layer (a) and underlying finer sediment (b) on an alternate bar along the Poudre River, Colorado, USA. Scale with inches (right) and centimeters (left) at upper left in (b).

threshold of hydraulic force that must be exceeded before the bed is mobilized, as well as increasing the hydraulic roughness of the bed. The coarse surface layer is characteristically one grain diameter in thickness and is both more coarse grained on average and better sorted than the underlying sediment. The underlying sediment has a grain-size distribution similar to that of the bed load in many gravel-bed rivers.

The coarse surface layer can be called pavement, armor, or a censored layer. Each of these terms is used differently by different investigators. *Pavement* is most commonly used for a coarse surface layer that is rarely disrupted. *Armor* most commonly refers to a coarse lag layer developed at waning flows that is regularly disrupted. *Censored layer* commonly refers to a layer that forms as finer matrix material is removed from around the surface framework particles as flow increases (Carling and Reader, 1982).

The mechanism by which the coarse surface layer forms, and the stability and frequency of mobilization of the layer, vary among channels and with time for a given channel. Coarse surface layers typically are well developed when local bed load supply from upstream is less than the ability of the flow to transport the bed load. This situation is common downstream from dams (Dietrich et al., 1989), and may develop with time following a pulse sediment input such as that from a volcanic eruption (Gran and Montgomery, 2005). Ephemeral channels typically have less developed coarse surface layers than perennial channels (Hassan et al., 2006). Coarse surface layers can also be spatially variable at the reach scale as a result of variations in shear stress and sediment dynamics associated with alternate bar topography, hydraulic resistance caused by instream wood, or variations in sediment supply.

Several studies suggest that the coarse portion of a surface layer moves down a river at nearly the same rate as the finer sediment in the layer. Coarse grains are intrinsically less mobile than finer grains, but the coarse surface layer might provide an equalizing mechanism by exposing proportionally more coarse grains to the flow. The coarse surface layer can thus be a mobile-bed phenomenon present under a range of flows, with only a few clasts of diverse sizes being entrained at any instant by even the peak flows (Andrews and Erman, 1986; Clayton and Pitlick, 2008).

Bed load in channels with coarse-grained substrate: characteristics of grain movements

In coarse-bed streams, grains move as individual particles or in discontinuous pulses. Pulses have been described as *sheets* in which migrating slugs of bed load 1 to 2 grain-diameters thick alternate between fine and coarse particles. Pulses of bed load can also take the form of *traction carpets*, which are highly concentrated bed load layers, or *streets*, which are longitudinally continuous tongues of bed load that do not span the full width of the channel. Pulses can also occur as *waves*, which are groups of low-amplitude bars migrating downstream in braided rivers, with plane bed transport across the upper bar surface and avalanching on downstream faces. Each of these forms of transport reflects the fact that coarse bed load is unlikely to be equally distributed in time and space along a reach of channel.

In channels with mixed grain sizes, rolling and sliding grains move episodically. Field observations suggest that rolling and sliding grains move less than 20% of the time during which flow conditions are sufficient for entrainment (Haschenburger, 2013), although periods of transport increase as rates of grain entrainment increase. Turbulent bursts cause much of the grain motion (Nelson et al., 1995).

Step length is the travel distance of a particle during continuous motion. The mean step length of grain motion appears to exceed 100 grain diameters, although step length increases with flow and decreases with increasing grain size and lower sphericity (Schmidt and Ergenzinger, 1992). In particular, step length varies only moderately for grains smaller than D_{50}, whereas step length of larger grains correlates inversely with grain size because the step length is limited by the energy required for transport of these larger grains (Church and Hassan, 1992). Grain collisions or areas of lower hydraulic force or rougher surface topography initiate periods of rest. The distributions of step lengths and rest periods have been described using gamma and exponential density functions, as well as a Poisson distribution.

Virtual velocity of grains quantifies the rate of downstream movement, including both step and rest periods during flows when entrainment can occur. This is known as *grain velocity*. Grain velocity, which is not particularly sensitive to flow velocity but does reflect grain characteristics such as density, can also be described with a gamma density function. Mean virtual velocity increases with flow, and has been documented at up to 100 m/h (Hassan et al., 1992).

The *path length* of moving bed load is the total downstream displacement of particles, typically through multiple steps, during a period of competent flow (Haschenburger, 2013). Local flow and bed conditions strongly influence the start and end of step lengths and path lengths. Longer duration flow can create longer path lengths when grains move over relatively smooth beds. Bed morphology such as bars or pools can limit path lengths for even longer duration flows, such that the modal path length correlates to spacing of pool—riffle–bar or step–pool units (Pyrce and Ashmore, 2003).

The number and size of grains entrained from the surface increase with flow and, when the entire surface is mobile, grains entrain from the subsurface. As the bed transitions from being immobile to partial mobility and then full mobility, the upper limit to partial mobility is *equal transport mobility*, where all grain sizes are transported in proportion to their presence on the bed (Parker and Toro-Escobar, 2002). This condition, which is relatively uncommon in coarse-grained channels, occurs when all surface grains are mobilized at flows exceeding entrainment thresholds by about four times (Wilcock and McArdell, 1997).

The distinction between partial and full mobility is also known as phase I and phase II transport (Andrews and Smith, 1992). *Phase I transport* starts when flow energy is just sufficient to rotate some of the gravel-sized particles out of their resting place, sending the particles rolling or bouncing downstream. As shear stress increases, more particles become mobile. Where a well-developed coarse surface layer is present, phase I involves transport of a finer fraction over a stable coarse bed. Sometimes, three phases of bed transport are distinguished. In this case, phase I is transport of a finer fraction over a stable coarse bed and phase II is partial and usually size-selective local entrainment (Ashworth and Ferguson, 1989; Warburton, 1992). Most of the bed is mobile during *phase II transport* (or phase III in a three-phase conceptualization). The two phases can be quantified using dimensionless shear stress, τ^*, where $\tau^* = \tau/g(\rho_s - \rho)D$: phase I occurs in the range $0.020 < \tau^* < 0.060$, and phase II when $\tau^* > 0.060$.

The nonlinear effects of sand on the transport of coarse particles in a bed with mixed grain sizes illustrate the complexities of bed load movement (Wilcock and Kenworthy, 2002). Sand can be preferentially transported at low discharges that do not move coarse sediment. As the amount of sand present increases, local sandy areas decrease the protrusion of large grains. Moderate amounts of sand can thus decrease the entrainment of large grains. Once a large grain is entrained, though, it can move faster over the relatively smooth sand bed and may move farther because sand fills the pore spaces between large grains and reduces resting places.

In addition to moving downstream, grains on the bed surface are also vertically exchanged with those in the subsurface through local scour and fill. Under some conditions, the bed is inactive and mobile grains pass over the bed. Finer grains can preferentially infiltrate, particularly if framework gravels on the bed dilate prior to movement (Allan and Frostick, 1999). (Dilation refers to a phenomenon in which grains lift and move apart from one another just before entrainment (Allan and Frostick, 1999).) More active beds facilitate grain exchange and sorting, although finer particles are still exchanged into the subsurface more quickly than coarser particles. Increased flow magnitude or duration increases the frequency and depth of vertical exchange (Wong et al., 2007), with exchange depths reaching up to 10 times the diameter of D_{90}, but more typically $2D_{90}$ (Haschenburger, 2013).

Entrainment and transport of alluvium along bedrock rivers likely varies along a continuum from an end-member with continuous alluvial cover, in which processes operate as described in Section 4.1.2, to a bedrock end-member in which sediment entrainment and transport are independent of size and more likely to reflect the location of local roughness and hydraulics (Goode and Wohl, 2010a) that stabilize alluvial sediment patches (Hodge et al., 2011).

Cold-region rivers have unique forms of sediment transport related to the formation and movement of frazil and anchor ice. *Frazil ice* is a collection of loose, randomly oriented, needle-shaped ice crystals that resembles slush. Frazil ice forms sporadically in open, turbulent, supercooled water, typically on clear nights when the air temperature drops to −6°C or lower. *Anchor ice* is submerged ice attached to the streambed. Anchor ice can form when large-scale turbulence mixes suspended ice crystals and frazil ice across the full depth of flow, allowing some of the ice to adhere to the bed or individual boulders. This is particularly common in riffles and can result in anchor ice forming in flows as deep as 20 m (Ettema and Daly, 2004). Under sufficiently cold temperatures and substantial flow turbulence, extensive areas of the streambed can become covered by anchor ice. Larger amounts of anchor ice typically form on coarser beds, which facilitate heat flux from a sub-bed zone that is 1–2 degrees above freezing to supercooled flow over the full depth above this zone. Anchor ice is less likely to form on fine, non-cohesive sediment, which is readily lifted by ice and therefore cannot hold a large accumulation of ice (Ettema and Daly, 2004).

Both frazil and anchor ice can be at least briefly attached to the bed and then float up with sediment frozen into the ice. Diurnal formation of these types of ice can result in repeated ice-rafting of sediment along a river during the cold season. Although observations indicate that sediment is entrained and rafted downstream by these forms of ice, there is little quantitative information on exactly how important this process is relative to other forms of sediment transport (Ettema and Daly, 2004). Much of the sediment entrained by the ice is stored in a seasonal ice cover until the cover breaks up during warmer weather.

Bed load in channels with coarse-grained substrate: controls on bed load dynamics

Spatial and temporal discontinuities in bed load movement have been explained in relation to several potential controls, each of which may dominate in different channels or as conditions within a channel change. Multiple controls can also influence bed load movement at a particular time and place.

Temporal variations in bed load can be distinguished as occurring at three scales.

1. Long-term variations reflect the rate of sediment supply to the channel (Vericat et al., 2006). Altered sediment supply from outside the

channel can occur as discrete sediment inputs of large volume, such as a landslide, that translate downstream as waves or disperse (Lisle, 2008; Sklar et al., 2009). Increased sediment can also enter the channel in prolonged, widespread inputs that produce responses such as an increase in gravel bars with a broad-scale wave-like form that move downstream over periods of tens to hundreds of years (Jacobson and Gran, 1999).

2. Short-term variations in bed load reflect temporary changes in sediment supply and the movement of bedforms, such as dunes (Gomez et al., 1989). Bedforms may not span the entire channel cross section, and time gaps can occur between passage of successive bedforms. During a single flood lasting hours to weeks, bed load can display hysteresis analogous to that of suspended sediment, with greater bed load transport rates during the rising stage. Discontinuities in bed load movement have also been attributed to longitudinal sediment sorting, which results in pulses of coarser or finer material moving downstream (Whiting et al., 1988). Lateral shifting of bed load streets, which typically do not span the entire channel width, can create spatial discontinuities in bed load and apparent temporal discontinuities at a point (Ergenzinger and de Jong, 2002). Sediment storage in secondary channels can fluctuate nonlinearly with changes in stage, resulting in discontinuities in supply to the main channel. Bed load discontinuities can reflect cross-sectional to reach-scale heterogeneity of bed structure and relief that influence both sediment supply and transport. In a channel with mixed grain sizes, for example, protruding clasts can create lee storage of finer sediment until fluctuating discharge alters the local hydraulics and the finer sediment is entrained (Figure S4.9) (Thompson, 2008).
3. Instantaneous variations in bed load reflect the inherently stochastic and probabilistic nature of the processes governing entrainment, as examined in Section 4.1.2. Typically, instantaneous variations increase with channel boundary roughness and increasing spread of bed grain sizes.

The frequency of the effective discharge for bed load decreases as the size of the bed sediment increases. In other words, coarser beds are mobile less frequently than finer beds. Rates of bed load transport increase rapidly and nonlinearly within a channel once sand becomes mobile, and can span up to seven orders of magnitude (Milhous, 1973).

The migration of readily mobilized bedforms can account for the majority of bed load movement in coarse-grained channels, as in sand-bed channels. Bedforms can also strongly influence the hydraulics that control sediment transport. In pool–riffle sequences, the pools have higher sediment transport during high flows, whereas riffles and alternate bars are sites of sediment deposition during high flows (Thompson et al., 1996). The steepened water-surface gradient over riffles and bars during low flows can promote dissection of the riffles and bars and the removal of finer particles that are then stored in pools until the next high flow. Clasts in pools of step–pool sequences are also preferentially entrained during the rising limb of a flood and deposited during the falling limb.

Instream wood can be a major source of cross-sectional to reach-scale heterogeneity of bed structure and relief, and can exert a particularly strong influence on bed load transport in channels with mixed grain sizes. By creating form roughness and obstructions, wood increases flow resistance and flow separation, leading to enhanced entrainment of material from portions of the bed and banks subject to scour because of current deflected by wood. Wood appears to be most effective, however, in promoting localized sediment storage (Figure S4.10). Many studies document preferential storage of bed material near individual wood pieces or logjams and greater overall storage, particularly of finer sediment, in channel segments with abundant wood (Piégay and Gurnell, 1997; Buffington and Montgomery, 1999; Montgomery et al., 1996, 2003b; Skalak and Pizzuto, 2010). Wood that breaks or is mobilized releases a pulse of bed load, and long-term removal of wood can change a depositional channel reach to a net sediment source (Brooks et al., 2003).

In addition to the physical processes acting to move bed load, aquatic organisms can displace or sort bed material and alter the susceptibility of bed sediment to entrainment and transport. Various species of fish create redds or nests for their eggs or larval fish. Where large numbers of fish such as salmon concentrate in a channel during the spawning season, their disruption of the bed can result in nearly half of the annual bed load yield (Hassan et al., 2008, 2011). Crayfish also disrupt the bed sediments in the course of moving about and burrowing into the bed, and these activities can significantly influence clast entrainment and transport during low flows (Statzner and Peltret, 2006).

The magnitude, frequency, and duration of bed load transport also influence aquatic biota by limiting the stability of the channel substrate and physically disturbing plants and animals via abrasion, dislodging, and displacing or burial. Even if the coarse surface layer of a mixed grain-size bed remains mostly stable, sand and finer gravel moving across the coarse surface can disturb benthic or bottom-dwelling organisms. Other factors being equal, sand-bed channels typically have lower abundance and diversity of benthic organisms than less mobile substrates, and less mobile features such as instream wood attached to the bank have a disproportionately large concentration of organisms in sand-bed channels (Minshall, 1984).

The greatest biological influence on bed load transport is undoubtedly human alterations of rivers. Changes to channel form through activities such as dredging, channelization, bank stabilization, and straightening influence the erodibility of bed and bank sediments and thus sediment supply. Sediment supply to the channel is also strongly influenced by human activities outside of the channel, as discussed in Supplemental Section 4.4 and Section 8.1.2.

Likely, the greatest alteration of suspended and bed load transport in many rivers is associated with flow regulation. Dams effectively trap all bed load and much of the suspended load moving down a river, leading to dramatically reduced sediment supplies downstream and a range of channel adjustments, from enhanced bank erosion to bed coarsening and incision. Globally, over 100 billion metric tons of sediment are now stored in dams, and sediment transport to the oceans has been reduced by an estimated 1.4 billion metric tons per year (out of 14 billion metric tons per year total) (Syvitski et al., 2005).

Estimating bed load flux

Bed load can be examined using a Lagrangian approach, which tracks grains along a river network, or an Eulerian approach, which describes grain passage at a fixed location such as a limited segment of channel (Doyle and Ensign, 2009). Using an Eulerian perspective, bed load flux can be estimated using three methods (Haschenburger, 2013).

1. Flux can be estimated from grain kinematics—the movement of grains as described by their displacement and velocity. This approach relies on coupling grain entrainment rate from a given bed area with the associated displacement length of the mobilized grains. Entrainment rate and displacement length are then extended to explicitly account for grain-size fractions (Wilcock, 1997; McEwan et al., 2004). Bed load transport rates can also be derived from the spatial concentration of bed load traveling at a given grain velocity. Or, transport rates based on the displacement of grains over some fixed time, together with the portion of the bed that is regularly mobilized, can be used to estimate flux.
2. Flux can be estimated from changes in river morphology that reflect bedform migration or bed aggradation and degradation (Trayler and Wohl, 2000; Haschenburger, 2006). This approach depends on being able to directly access the bed using survey instruments, or indirectly image the bed with fathometers or other remote sensing, at time intervals and spatial resolution relevant to bed load flux.
3. Bed load transport rates can be derived from (semi) empirical equations, typically based on laboratory experiments. Starting with Shields' flume experiments in sand-bed channels

(Section 4.1.2), investigators initially developed *bulk models of bed load flux* with equations based on one representative grain size. Most of these models were variations on Shields' formula and assumed that transport is proportional to a bulk flow parameter such as shear stress or stream power. An example of a bed load transport equation that follows this approach and remains widely used is the Meyer-Peter and Mueller (1948) equation, which can be written in dimensionless form as

$$\frac{g_b}{\left[\left(\frac{\rho_s - \rho}{\rho}\right) g D^3\right]^{\frac{1}{2}}} = 8\left(\theta' - \theta_c\right)^{\frac{3}{2}} \quad (4.14)$$

where ρ is water density, ρ_s is sediment density, g is acceleration due to gravity, D is the size of the bed material, g_b is the transport rate (volumetric transport rate per unit width, in m^2/s), θ' is the dimensionless grain stress, and θ_c is the critical dimensionless shear stress (McLean et al., 1996). In the original form of the equation, θ_c was set equal to 0.047 (Robert, 2003). Actual transport rate is assumed to equal theoretical transport capacity in this equation, although rate is likely to be lower than capacity in coarse-grained channels because of limited sediment supply.

Starting in the 1980s, size-specific formulae were developed for bed load flux. This approach assumes that different grain sizes move independently of each other, although movement or stationarity by grains of a given size can influence the movement of other grains via effects such as shielding. An example of this approach is the *two-fraction transport model* proposed by Wilcock and Kenworthy (2002). The two fractions of sand and gravel are described by the dimensionless transport rate W_i^* via

$$W_i^* = \frac{(s-1)\, g q_{bi}}{F_i u_*^3} \quad (4.15)$$

where s is the ratio of sediment to water density, g is gravity, q_{bi} is volumetric transport rate per unit width of size i, u_* is shear velocity [$u_* = (\tau/\rho)^{0.5}$, where τ is bed shear stress], F_i is proportion of fraction i on the bed surface, and the subscript i represents either the sand or gravel fraction.

More recent research focuses on *grain–grain interactions* during transport (Frey and Church, 2012). Experiments with sand and gravel tracers reveal diffusion of the tracers, statistically defined as the spreading of particles downstream. Diffusion reflects the fact that individual grains move at different rates as a result of the complexity of grain–grain and grain–fluid interactions (Martin et al., 2012).

The progression of differing approaches through time reflects the application of increasingly sophisticated physics to the problem of understanding and predicting bed load flux, particularly in channels of mixed grain size. Changes in flow and bed load transport rate typically do not correspond well in space or time in these channels, for the many reasons outlined previously, making prediction of bed load transport extremely difficult.

Whatever their form, most bed load transport equations do not accurately predict fluxes in channels with coarse or mixed grain sizes. The discrepancy reflects at least three factors (Haschenburger, 2013). First, the equations require a specific channel response based on defined hydraulic conditions. In other words, the equations require equilibrium transport conditions, which do not always occur in coarse-grained channels. Second, the equations do not integrate the spatial and temporal complexities that influence entrainment and transport, such as spatial and temporal heterogeneities in bed roughness, which can be substantial in coarse-grained channels. Finally, the equations rely on input data that are typically difficult to characterize, such as spatial heterogeneity of the bed.

Prediction of bed load flux for specific rivers can be improved by collecting bed load data. Supplemental Section 4.2.3 provides details on measuring bed load.

4.3 Bedforms

Bedforms are bed undulations that result from sediment transport. Bedforms in sand-bed channels are smaller than channel-scale bar forms. Sand bedforms are sometimes distinguished as:

- *microforms* such as ripples for which the occurrence and geometry are controlled by boundary-layer characteristics including bed grain size;
- *mesoforms* such as dunes that are controlled by boundary layer thickness or flow depth; and
- *macroforms* such as bars that have lengths of the same order as channel width and heights comparable to the flow producing the form (Bridge, 2003).

Bridge (2003) also distinguished between a bedform—a single geometrical element such as a ripple, and a *bed configuration*—the assemblage of bedforms of a given type that occurs in a particular bed area at a given time. Bedforms will be used here in a more general sense that includes individual bedforms and the bed configuration.

Explaining bedforms involves considering both sediment transport dynamics and channel geometry. Migration of bedforms creates bed load flux, and alters the conditions for mobility of sediment not within a bedform. In the presence of bedforms, a significant proportion of the total stress is caused by form drag on the bedform and this component of the total stress is not effective in moving sediment. Bedforms therefore reduce bed load transport rates relative to a flow with the same mean velocity that does not include bedforms.

The presence and characteristics of bedforms are interconnected with channel geometry to the extent that the occurrence of certain types of bedforms reflects not only channel geometry variables such as flow width and depth or bed gradient, but also influences these variables. Step–pool sequences, for example, form within a restricted range of bed gradient. Steps and pools also alter flow path length by creating a stepped rather than uniform longitudinal profile.

As noted in the introduction to this chapter, the subdivisions within this section recognize an important difference in bed mobility. At one extreme are sand-bed channels in which readily mobile bed substrate results in relatively high bed load fluxes and quickly changing bedforms. At the other extreme are cohesive-bed channels in which the bed deforms only by erosion. Between these end-members are coarse-grained channels in which a larger threshold of flow energy (relative to sand-bed channels) must be exceeded before the bed material begins to move and bedforms are created or mobilized.

4.3.1 Readily mobile bedforms

Readily mobile bedforms are primarily those in sand-bed channels, which occur in the broad classes of small-scale ripples and larger-scale dunes, both of which have their long axes perpendicular to downstream flow. Other types of bedforms that develop under specific hydraulic conditions and are rarer include upper-stage plane beds and antidunes (Figure 3.13).

Understanding of bedforms in sand-bed channels is particularly important for at least three reasons. First, bedforms are the most important source of flow resistance at the local scale, creating two to three times the resistance caused by grains. Second, bedform initiation, growth, and migration dominate sediment dynamics in sand-bed channels, both by serving as a basic transport process and by affecting the near-bed flow field and distribution of hydraulic force and turbulence. Finally, bedforms leave characteristic depositional records, the interpretation of which is fundamental to understanding past river environments (Venditti, 2013).

A typical sequence of bedforms evolves as velocity increases (Figure 3.13). At very low velocities, sand grains on a bed will not move. As velocity increases, the lift and drag forces exceed the submerged grain weight and other resisting forces. The bed begins to develop millimeter-scale grooves parallel to the flow and spaced at the scale of the high and low speed streaks that are ubiquitous in shear flow. This is a *lower-regime plane bed*, which can persist for long periods of time or develop into bedforms as a result of either defects or instantaneous initiation (Venditti et al., 2005). Defects such as a small depression or protrusion into the flow create flow separation and local shear stresses able to move sediment while the average shear stress remains below the threshold of motion. At higher shear stress and general sediment transport, bedforms appear over the entire bed nearly instantaneously. This presumably reflects instability in the form of turbulence and velocity fluctuations at the water–sediment interface that rapidly increase the extent of entrainment and grain–grain interactions, allowing grains to start collecting into bedforms.

The stages in Figure 3.13 up through dunes are *lower flow regime bedforms*. These typically exist in subcritical flow and have relatively small bed load transport rates. The size of lower-regime bedforms varies widely between diverse rivers, from a few centimeters to several meters in height and a few tens of centimeters to as much as a kilometer in length. Bedform height, H, and length, L, are empirically related (Flemming, 1988) as

$$H = 0.677L^{0.8098} \tag{4.16}$$

The existence of two subpopulations in a plot of H versus L is used as evidence that ripples and dunes are distinct forms (Ashley, 1990). Ripples do not affect the water surface and form only in sediment where $D < \sim 0.7$ mm. When dunes form, waves develop on the water surface that are out of phase with the dune troughs and crests.

Dunes form in sediment from fine sand to gravel size. The ability to form dunes thus reflects flow magnitude, rather than grain size, as evidenced by dunes formed of boulder- to gravel-sized particles during late Pleistocene outburst floods (Baker, 1978). Dunes are typically less steep than ripples, although slope angles for the two types of bedforms overlap.

Lower-regime bedforms can be low-angle symmetric (downstream slope ~10 degrees) or angle of repose asymmetric with an upstream slope averaging 2–6 degrees and downstream slope ~30 degrees. Symmetric forms are common in large rivers and estuaries, whereas asymmetric forms seem to be more common in small channels, although it remains unclear how mechanics of formation differ between the two types.

The planimetric morphology of lower-regime bedforms can be distinguished as being two or three dimensional. Two-dimensional (2D) bedforms have fairly regular spacing, heights and lengths, and straight or slightly sinuous crestlines oriented perpendicular to the mean flow lines. Three-dimensional (3D) bedforms have irregular spacing, heights and lengths with highly sinuous or discontinuous crestlines (Venditti, 2013; Figures 4.7 and S4.11). As with symmetry, differences in the mechanics of formation of 2D versus 3D bedforms remain under discussion, although 3D forms may reflect persistent, unidirectional flows, whereas 2D bedforms reflect variable flows. Lower-regime bedforms can also be superimposed on one another, as when ripples and dunes are superimposed on bars, or ripples are superimposed on dunes.

As the Froude number approaches 1, dunes are increasingly likely to wash out to a plane bed, which is the start of the *upper flow regime bedforms* (plane beds can be present at Froude numbers well below 1). These bedforms exist in supercritical flow and have relatively large bed load fluxes and small flow resistance. Upper-stage plane beds have low-relief bed waves only a few millimeters in height. These bed waves migrate downstream beneath a water surface with waves that are out-of-phase with the bed waves. At the high velocity end of upper flow regime bedforms, antidunes occur in all grain sizes and the water surface waves are in-phase with the bed sediment waves. Antidunes are upstream-moving undulations that typically grow to an unstable steepness, break, and then reform (Southard, 1991). Antidunes can form from dunes or upper-stage plane beds as velocity increases (Venditti, 2013).

Bedforms can also be described using phase diagrams. Figure 4.7 illustrates the distribution of bedform types in relation to various hydraulic and sediment variables, including fall velocity, stream power, Froude number, and dimensionless excess shear stress (Venditti, 2013). Examining bedforms in the context of a phase diagram emphasizes the relation of individual types of bedforms to a continuum defined by hydraulic and sedimentary variables (Southward and Boguchwal, 1990; Southard, 1991) (Figure S4.12).

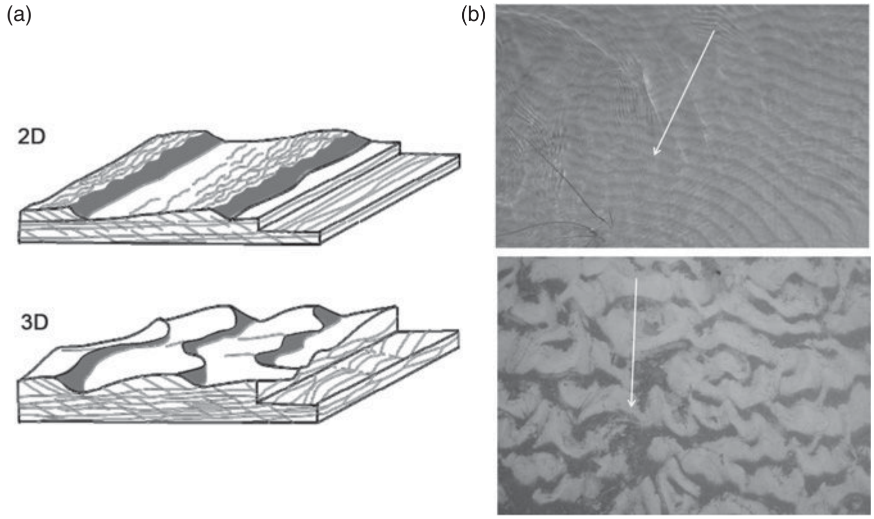

Figure 4.7 (a) Schematic illustration of the planimetric morphology of two-dimensional (2D) sand bedforms with planar cross-stratification and three-dimensional (3D) sand bedforms with trough cross-stratification (from Venditti, 2013, Figure 7). (b) Plan view photographs of 2D (upper photo) and 3D (lower photo) ripples, both in the shallows along the edge of the Niobrara River of Nebraska, USA. Sedge stem in upper photo gives some indication of scale: both sets of ripples are a few centimeters tall. Primary flow direction is indicated by a white arrow in each photograph. Dark material in lower photograph is fine-grained organic matter temporarily deposited in the lee of each ripple.

Most examinations of flow over sand bedforms have focused on asymmetric, 2D dunes. Flow accelerates and converges over the upstream side of these dunes, then separates at the dune crest. The zone of flow separation is associated with a turbulent wake and shear layer originating at the dune crest and extending and expanding downstream. Flow reattaches at a distance downstream about four to six times the bedform height. An internal boundary layer grows from the reattachment point downstream beneath the wake toward the crest. An outer, overlying wake region is present above the level of flow separation and reattachment (Venditti, 2013) (Figure 4.8a).

Flow over symmetric dunes includes a well-defined region of decelerated flow in the lee of the bedform, a shear layer between the main flow and the decelerated flow, and turbulent eddies generated along this shear layer that dominate the macroturbulent flow structure (Venditti, 2013) (Figure 4.8b). Symmetric dunes have less intense mixing along the shear layer in the dune lee, and only intermittent flow separation.

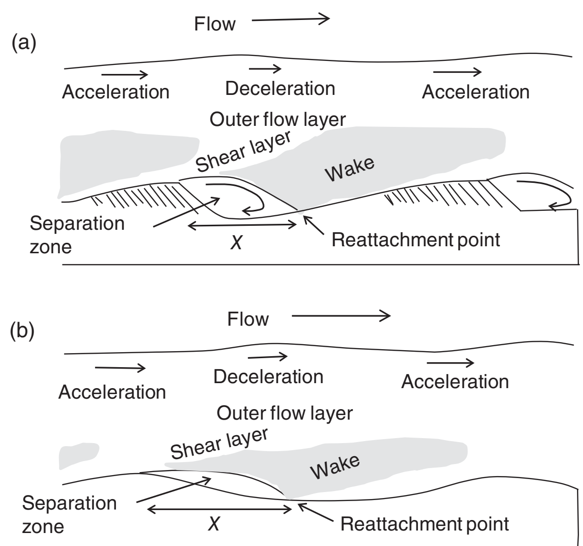

Figure 4.8 Hydraulics over sand bedforms. (a) Two-dimensional asymmetric dunes at the angle of repose and (b) low-angle asymmetric dunes. X is the reattachment length (from Venditti, 2013, Figure 9a and 9b).

All forms of coherent flow structures and turbulence over asymmetric and symmetric dunes effectively increase the resistance of a channel with such bedforms and influence sediment entrainment and mobility. Symmetric dunes have lower flow resistance, however, than asymmetric dunes.

The structure of turbulence over ripples and dunes controls the flow resistance. The flow structure over dunes is dominated by depth-scale turbulent eddies, known as *macroturbulence* or *kolks* when they are in the water column and *boils* when they disturb the water surface (Venditti, 2013). Kolks and boils likely originate from the combined effects of boundary layer bursts, instabilities along the shear layer, and vortex shedding.

Individual bedforms continue to change their geometry under constant flow conditions, as well as under changing flow. Bedform changes typically lag changes in flow, and thus display *bedform hysteresis* that can be minutes to hours for ripples, and as much as days to months for dunes (Venditti, 2013).

In addition to changing their geometry, bedforms migrate with time, thus contributing to sediment transport. Asymmetric bedforms migrate by erosion on the upstream side and deposition on the downstream side via avalanching down a steep slipface. Over symmetric dunes, migration can also involve large amounts of sediment entrained into suspension on the upstream side of the dune and deposited from suspension on the downstream side.

Migration creates inclined packets of sediment that can be preserved as cross-strata in the sedimentary or rock record. Ripples typically create cross-strata millimeters thick, and dunes can create cross-strata millimeters to centimeters thick. Planar cross-strata formed by 2D bedforms (Figure S4.13) are parallel to the bedform slipface. These beds form over a planar erosional surface scoured by a laterally continuous recirculation cell as the bedform trough passes downstream. Trough cross-strata formed by 3D bedforms (Figure S4.13) accumulate over a curved erosional surface resulting from flow separation in the lee of a saddle-shaped bedform crest (Bridge, 2003).

4.3.2 Infrequently mobile bedforms

Infrequently mobile bedforms are primarily those occurring in gravel-bed channels. The cobbles and coarser sediment in these channels have a higher entrainment threshold that may be exceeded much less than half of the time during an average flow year, although minor adjustments associated with individual grain movements or the mobility of the finer fraction of the bed sediment can occur over a wide range of flows. Lower-regime plane beds and dunes, as well as upper-stage plane beds and antidunes, can occur in gravel-bed channels (Bridge, 2003).

As with the more readily mobilized sand-bed bedforms, infrequently mobilized bedforms in coarser-grained channels both respond to and strongly influence flow resistance, the distribution of velocity, turbulence, and other hydraulic variables, and sediment transport. The interactions among bedforms, flow, and sediment in transport within gravel-bed channels typically occur over larger spatial scales and longer timescales than analogous interactions in sand-bed channels, but exhibit similar levels of complexity. The primary bedforms treated here are particle clusters, transverse ribs, pool–riffle sequences, step–pool sequences, and bars.

Particle clusters

Particle clusters are closely nested groups of clasts aligned parallel to flow, typically 0.1 to 0.2 m in length in the downstream direction, and about twice as long as they are wide (Brayshaw, 1984) (Figure S4.14). These features are also known as pebble clusters, imbricate clusters, cluster bedforms, and microforms. Particle clusters can take a variety of shapes, including comet shaped, triangular, rhomboid, or diamond. Typically, an obstacle clast anchors an upstream-side accumulation of imbricated particles with a tail on the downstream side (Papanicolaou et al., 2003).

Particle clusters are most common in gravel-bed channels with low rates of bed load transport and a stable coarse surface layer, and usually occur in riffle, alternate bar, or plane-bed sections of the channel. Although more than one mechanism of formation may exist, clusters appear to form during the recessional limb of a flood as bed load is deposited around large, protruding clasts (Brayshaw, 1984). Clusters create abrupt roughness transitions along the bed and increase average boundary roughness (Hassan and Reid, 1990). This leads to localized flow separation and turbulence (Lacey et al., 2007). By creating relatively stable points on the streambed, clusters also delay incipient motion and limit the availability of bed material for transport (Brayshaw, 1984).

Transverse ribs

Transverse ribs, like particle clusters, are microtopographic features that can be difficult to see when looking at a streambed. Ribs are formed by a series of regularly spaced pebble, cobble, or boulder ridges perpendicular to the flow. The spacing between ribs is proportional to the size of the largest clasts in the rib crest (Robert, 2003). Spacing between ribs is typically of the order of decimeters to meters, and heights are one to two clast diameters (Bridge, 2003). Ribs appear to be most common on bars in gravel-bed rivers.

Transverse ribs likely represent the crests of antidunes (Koster, 1978). Water-surface waves commonly break upstream of antidune crests, creating temporary hydraulic jumps that could promote coarse sediment deposition. As with other bedforms, the presence of transverse ribs influences boundary resistance and sediment entrainment and transport.

Steep alluvial channel bedforms

The classification of reach-scale channel morphology proposed by Montgomery and Buffington (1997) for steep, relatively laterally confined gravel-bed channels is largely based on dominant bedforms in these channels (Figure 4.9). Channels with the

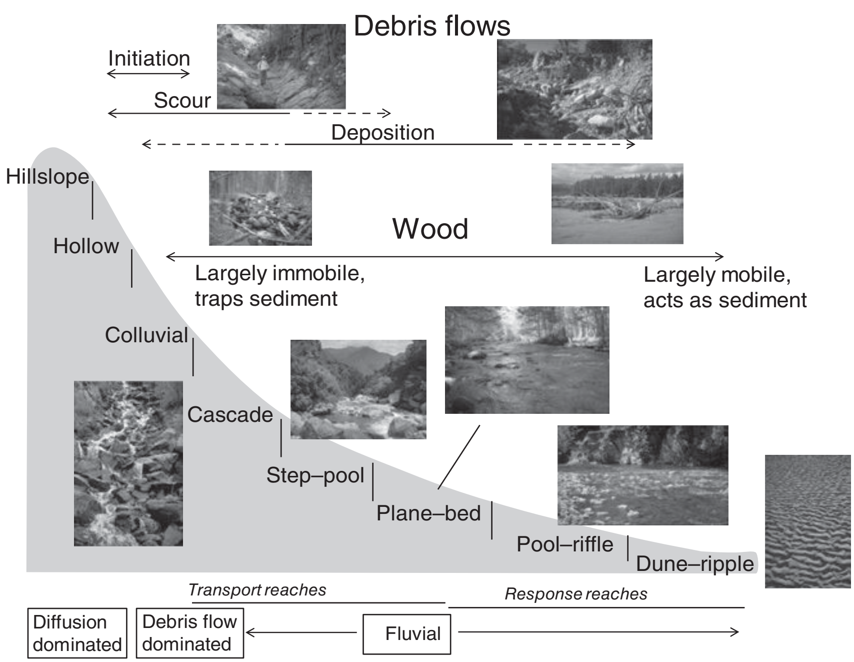

Figure 4.9 The Montgomery and Buffington (1997) channel classification based on dominant bedforms. This schematic illustration of downstream trends includes changes in reach-scale gradient (light gray shading), channel type (cascade/step–pool/plane-bed/pool–riffle/dune–ripple), indication of transport versus response reaches, dominant sediment transport process (hillslope diffusion/debris flow/fluvial), and the behavior of debris flows and instream wood (from Montgomery and Buffington, 1997, Figure 4).

highest bed gradients are typically *cascade channels* with disorganized, very coarse bed material and small pools that partially span the channel and are spaced less than a channel width apart. In these channel segments, flow may not be competent to organize the very coarse-grained bed material into bedforms.

At gradients of 0.03–0.10 m/m, step–pool channels become particularly common. Although substrate mobility typically remains limited in these channel segments, flow is competent to create regularly spaced bedforms and bed variability in elevation and grain size, primarily in the downstream direction.

Plane-bed channels lack well-defined, rhythmically occurring bedforms and are most common at gradients of 0.01 to 0.03. The relatively planar bed formed in coarse clasts or bedrock can persist through conditions of bed stability and mobility. Despite the similar name, coarse-grained plane-bed channels are much less mobile than plane-bed sand channels.

Alternating pools and riffles characterize channels with gradients less than 0.015 m/m, which exhibit consistent patterns of bed variability both downstream and across the channel. Pool–riffle sequences transition at lower gradients to *dune–ripple channels* with sand beds and readily mobile bedforms.

The progression with decreasing reach-scale gradient from cascade to dune–ripple channels corresponds to decreasing average bed grain size and increasing bed mobility. This represents a general downstream trend that can be interrupted or reversed. Mountain rivers are commonly longitudinally segmented and alternate repeatedly downstream between high and low gradient reaches (Wohl, 2010), so that cascade segments can be downstream from pool–riffle segments, and vice versa. Step–pool and pool–riffle channels are particularly common and well-studied channel types within this progression.

Step–pool channels

The dominant bedforms in *step–pool channel* segments are channel-spanning steps of clast, wood, or bedrock that alternate downstream with a plunge pool at the base of each step (Chin and Wohl, 2005; Zimmermann, 2013) (Figures 4.10 and S4.18). Step–pool sequences are most common where relatively immobile large clasts or wood trap sediment in a wedge tapering upstream, with flow plunging over the immobile obstacle to scour a plunge pool in the bed downstream. Although steps likely form via more than one mechanism, those in mobile sediments (rather than bedrock) appear to result from the interlocking of large, *keystone* clasts

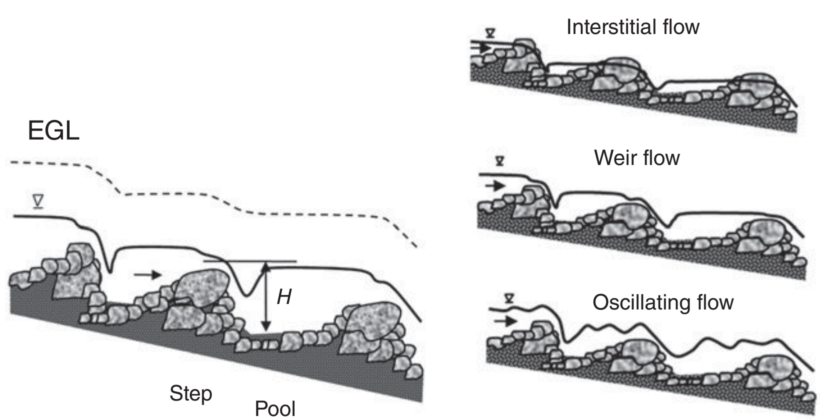

Figure 4.10 Longitudinal view of the components and hydraulics associated with step–pool bedforms. *H* is bedform amplitude: here, the step height. EGL is energy grade line. Interstitial, weir, and oscillating flows represent flow regimes under progressively increasing discharge. Flow is from left to right. (From Dust and Wohl, 2012a, Figure 5b; Dust and Wohl, 2012b, Figure 2.)

(Zimmermann and Church, 2001) under conditions of limited sediment supply, leading to a *jammed state* (Zimmermann et al., 2010).

Individual steps can be destroyed by erosion of the keystones and subsequent mobilization of the step (Lenzi, 2001). Step–pool sequences can give way to braided or plane-bed channels during extremely large discharges that mobilize the entire bed and greatly enhance the sediment supply. Such mobilization occurs anywhere from annually in some channels, to intervals of 50 years or longer in other channels (Wohl, 2010). As the mobilizing flood recedes and sediment supply once again declines, steps and pools gradually reform during subsequent, smaller flows. These smaller flows can mobilize the finer sediments exposed at the surface, but not necessarily the keystone clasts. An example comes from the 5 km^2 Rio Cordon catchment in Italy, where floods with recurrence intervals of 30–50 years destroy step–pool sequences and move individual clasts approximately twice the downstream distance of movement during smaller floods with recurrence intervals of 1–5 years. Selective entrainment and transport during the smaller floods gradually reform step–pool sequences (Lenzi, 2001, 2004).

Steps and pools create extremely large values of flow resistance relative to lower gradient channels (Curran and Wohl, 2003). Steps are typically defined by height, H, and downstream spacing, L, and many step–pool sequences have a ratio of $H/L/S$ (S is bed gradient) between 1 and 2. Bedforms with this ratio have been interpreted to maximize flow resistance and channel stability (Abrahams et al., 1995). Wood incorporated into clast steps can increase step height, backwater effects, and dissipation of flow energy (Wilcox et al., 2011). The large grain and form resistance of step–pool channels also creates high turbulence intensities, particularly in pools and at high discharges (Wilcox and Wohl, 2007).

Lower discharges in step–pool sequences can create interstitial flow, with water primarily passing through gaps between the step-forming clasts. With increasing discharge, free falls over steps create nappe flow or weir flow, with a hydraulic jump below the step. At the highest discharges, the hydraulic jump disappears and an oscillating or skimming flow regime develops in which the water flows as a coherent stream and recirculating vortices occur at the base of each step (Chanson, 1996; Dust and Wohl, 2012a) (Figure 4.10). Flow resistance decreases significantly when flow transitions from nappe to skimming conditions (Comiti et al., 2009).

Values of resistance coefficients such as Manning's n are very difficult to estimate in step–pool channels, which have non-uniform flow. An alternative approach to estimating discharge is to apply the equation developed for broad-crested weirs, under the assumption that a step approximates such a weir

$$Q = C^{*} g^{0.5} W h^{3/2} \tag{4.17}$$

where C^{*} is a dimensionless discharge coefficient, W is the crest width, g is the acceleration of gravity and h is the upstream flow depth above the step crest (Dust and Wohl, 2012a).

Bed load transport is extremely spatially and temporally variable and very difficult to quantify or predict in step–pool channels. Yager et al. (2012) proposed a modified version of the Parker (1990) bed load equation to include the resistance associated with steps and selective transport of relatively mobile sediment using a range of hiding functions.

Particles in pools are preferentially entrained and transported longer distances. After placing tracer clasts in a small step–pool stream in Germany, Schmidt and Ergenzinger (1992) found that all the tracers placed in pools were entrained during lower discharges than those in steps. All of the pool particles were transported, whereas 57%–76% of the particles placed in other sites on the bed were transported. Pool particles also had greater transport lengths. Particles finer than or equal to D_{40} of the bed surface in step–pool channels do not exhibit substantial differences in travel distance during floods, whereas particles larger than the bed surface D_{84} have very limited mobility (Lenzi, 2004).

Step–pool channels are known as *transport reaches* (Montgomery and Buffington, 1997) because they are relatively insensitive to changes in water and sediment supply. Field studies support the idea that, when water and sediment supplied to a river network change, step–pool bedforms are less likely to alter their dimensions than bedforms present at lower channel gradients (Ryan, 1997; Wohl and Dust, 2012). The resistance to change likely reflects the combined effects of high values of boundary irregularity and flow resistance that effectively dissipate flow energy, and very coarse, relatively interlocked clasts in steps.

Pool–riffle channels

Pool–riffle bedform sequences are analogous to meandering in the vertical dimension (Keller and Melhorn, 1978) in that regularly spaced deeps with typically finer sediment (pools) and coarser-grained shallows (riffles) create an undulating longitudinal profile at the reach scale (Figures 4.11 and S4.19). Riffles are usually wider and shallower at all stages of flow than pools.

Pool and riffle bedforms occur in alluvium and in bedrock, and are likely formed and maintained by diverse mechanisms. The amplitude and wavelength of these bedforms is typically described in terms of average downstream spacing of pools, which has been related to channel width (Keller and Melhorn, 1978). Pools are commonly described as being spaced at approximately five to seven times the average channel width. Despite this commonly cited rule, the relation between pool spacing and channel width is in reality quite variable. Pool spacing varies with substrate, such that pools tend to be more widely spaced in more resistant substrate such as bedrock (Roy and Abrahams, 1980; Wohl and Legleiter, 2003). Pool spacing also varies with valley width and associated flow convergence. Working on the Yuba River of California, USA, White et al. (2010) found that riffles persisted in locally wide areas of the valley and pools were associated with long valley constrictions over a period of 20 years, despite rapid channel incision and planform change in response to frequent floods.

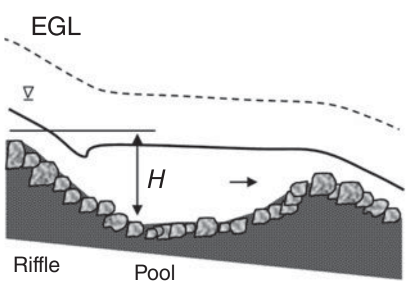

Figure 4.11 Longitudinal view of the components and water surface associated with pool–riffle bedforms. *H* is bedform amplitude: here, the elevation difference between the riffle crest and the next pool thalweg. EGL is energy grade line, shown here for relatively low flow conditions. During high flow, the EGL over riffles flattens, whereas that over pools grows steeper. Flow is from left to right. (From Dust and Wohl, 2012b, Figure 2.)

Pool spacing also varies with the longitudinal spacing of obstructions such as wood (Montgomery et al., 1995; Buffington et al., 2002) or large, immobile boulders (Lisle, 1986; Thompson, 2001), which typically create localized scour and pool formation. Pool–riffle sequences associated with local scour and deposition around an obstruction are sometimes referred to as *forced pool–riffle sequences* (Montgomery et al., 1995). Obstructions pond water upstream and create flow convergence, higher water-surface slopes, and higher flow velocity through the constricted zone. This creates localized scour that leads to pools. As the flow passes downstream beyond the obstruction, flow divergence and deceleration along the downstream end of the pool center facilitate sediment deposition, leading to formation of a riffle (Lisle, 1986; Thompson et al., 1998; Thompson, 2013).

Pools and riffles formed in relatively well-sorted alluvium with few obstructions are dominated by feedbacks between hydraulics and sediment transport, and are more likely to exhibit pool spacing at five to seven times channel width. Pools and riffles formed in poorly sorted, coarse-grained alluvium with abundant obstructions are also strongly influenced by hydraulic–sediment interactions, but local variations in channel boundary configuration exert a much stronger influence on bedform characteristics.

The locations of freely formed (rather than forced) pools and riffles have been explained in terms of at least two mechanisms of hydraulics and bed load movement. In some channels, kinematic waves of bed load appear to create bars and riffles where the wave stops moving during the falling limb or where transport capacity locally declines (Langbein and Leopold, 1968). Accelerating flow converging downstream from the bar then scours a pool. Another scenario involves persistent point sources of coarse sediment, such as tributary junctions, that create riffles (Webb et al., 1987). The riffles create accelerated flow that leads to pool scour downstream (Dolan et al., 1978). In each of these explanations, creation of an initial bed irregularity in the form of either a pool or a riffle then perturbs hydraulics and bed load transport sufficiently to initiate downstream bed undulations that lead to successive pools and riffles (Clifford, 1993). Average longitudinal pool spacing likely reflects a minimum distance related to the backwater and turbulent conditions needed for pool formation (Thompson, 2012).

The *velocity-reversal hypothesis* has been used to explain the maintenance of pools and riffles (Gilbert, 1914; Keller, 1971). Field observations indicate that the near-bed velocity increases more rapidly with discharge in a pool than in a riffle, so that flow in pools is more competent at high stage than flow over riffles. This explains the common observation that pools scour at high flow and fill at low flow, whereas riffles are depositional sites at high flow. The velocity reversal has been difficult to explain, however, given the tendency of pools to have greater cross-sectional area (and therefore presumably lower velocity) than riffles. Hydraulic modeling of the River Severn in England indicated that channels in which pools are hydraulically rougher than riffles during high flow, and riffles are substantially wider than pools, are more likely to exhibit velocity reversal (Carling and Wood, 1994).

Numerous field observations and flume experiments illustrate the difficulty of demonstrating a velocity reversal, at least in part because of the difficulty of measuring bed velocity during high discharges. Cross-sectionally averaged velocity does not exhibit velocity reversal, but the presence of strong eddy flow along the margins of pools during high discharges permits the formation of a central jet of high velocity flow that does exhibit velocity reversal with respect to riffle velocity (Thompson et al., 1998, 1999) (Figure S4.20).

Bed load in pool–riffle channels tends to move during competent flows in spatially discrete steps that are strongly influenced by pool and riffle spacing (Thompson et al., 1996; Pyrce and Ashmore, 2003). Clasts entrained from a pool during high flow are deposited on the next riffle downstream. Clasts on a riffle can be moved into the next pool downstream during waning or low flows as progressively steeper water-surface profiles over riffles at lower stages result in dissection of the riffle (Harvey et al., 1993).

In contrast to step–pool sequences, pool–riffle channel segments are *response reaches* (Montgomery and Buffington, 1997). Because pool–riffle channels are more likely to be transport limited with respect to sediment, bedform dimensions and bed grain-size distributions of pool–riffle channels are more likely to change in response to altered water and sediment discharge than in supply-limited step–pool channels. Increased sediment load typically causes preferential pool filling. Filling pools reduces form roughness and creates a more uniform reach-scale bed gradient and flow depth that enhance the ability of moderate flows to transport bed load (Lisle, 1982). Decreased sediment load and/or increased discharge can lead to enhanced pool scour and bank erosion (Wohl and Dust, 2012). The bedform amplitude and wavelength of pool–riffle sequences can thus adjust to variations in flow and sediment supply.

An important implication of the distinction between transport and response reaches is that diverse types of infrequently mobile bedforms are not likely to respond uniformly to changes in water and sediment yield associated with changes in climate, land cover, or other external controls. If timber harvest within a catchment increases sediment yield, for example, bedform dimensions in the lower gradient response reaches will change disproportionately, while those in the higher gradient transport reaches may remain relatively unaffected.

River reaches with different bedforms exhibit differences in *sensitivity* (the ability to react to a stimulus) and *resilience* (the ability of a river to return to initial conditions following a disturbance such as a flood or debris flow) (Brunsden and Thornes, 1979). These reach-scale differences within a river network have important implications for channel stability, aquatic and riparian habitat, water quality, and river management (Montgomery and Buffington, 1997; Wohl et al., 2007). Sensitive channel segments may be less stable, for example, whereas resilient segments may be more stable.

By influencing flow velocity and depth, substrate size and stability, dissolved oxygen, hyporheic exchange, and solute fluxes, infrequently mobile bedforms also influence habitat and the distribution of aquatic biota. In mountainous headwater channels, for example, pool–riffle channel segments typically have greater pool volume and fish habitat than step–pool or cascade channels (Moir et al., 2004).

Bars

Diverse types of bars occur in rivers, including berms, alternate bars in straight channels, point bars in sinuous channels, transverse bars that form riffles or rapids, braid bars or mid-channel bars in braided channels, and bars at tributary–main channel junctions. *Bars* have lengths comparable to channel width. Bars form when flow energy is sufficient to transport bed material and simultaneously scour other portions of the channel bed, such as pools. As with other bedforms, bars are both influenced by boundary resistance, hydraulic forces, and sediment dynamics, but also influence roughness, hydraulics, and sediment dynamics.

Berms are accretionary features that form at sites of energy loss associated with hydraulic jumps or the shear zone of flow separation (Carling, 1995) (Figure S3.16). Berms can form downstream from lateral channel constrictions or at the downstream end of a plunge pool below a vertical step such as a waterfall. Berms can be relatively small features that are the height and lateral thickness of a single cobble or boulder, or can be more than a meter tall and wide. Berms tend to be persistent features that form during high discharges and remain during subsequent lower discharges.

Alternate bars occur on alternating sides of the channel in a downstream progression. Alternate bars can form in mixed grain-size channels when concentrations of coarse clasts at the downstream end of bed load pulses aggrade along portions of the channel with high flow roughness (Lisle et al., 1991). Alternate bars are typically asymmetrical longitudinally and generally migrate in the downstream direction (Bridge, 2003) as particles being transported along the top of the bar reach the downstream edge and cascade down the bar front. These bars are discussed more in Section 5.2 as part of the treatment of meandering and braided channels.

Point bars form along the inside edge of meander bends, where fine grains are swept inward over the point bar and coarse grains are routed outward toward the pool (Clayton and Pitlick, 2007). Depending on the channel substrate and sediment supply, point bars can be formed of sand to boulder-size grains, and are scaled to channel width. Point bars are essentially alternate bars in sinuous channels.

Point bars migrate and adjust in size and geometry as sediment supply and the distribution of hydraulic forces fluctuate with discharge, and as meander geometry changes through time. An extreme example comes from the highly sinuous Fall River in Colorado, USA, which received a huge influx of sediment following a damburst flood in the headwaters. The flood waters eroded large volumes of glacial outwash sediment from the valley bottom en route to the meandering reach, where the resulting sediment supply increased by about 1000 relative to normal sediment influx (Anthony and Harvey, 1991). During the 4 years following the damburst, point bars aggraded to the bankfull water surface during higher flows, then eroded laterally and vertically during periods of lower flow when sediment supply from upstream decreased.

Point bars are also discussed more in Section 5.2.

4.3.3 Bedforms in cohesive sediments

Characteristic repetitive bed undulations occur in streambeds with sufficient silt- and clay-sized particles to act as cohesive materials. These bedforms are strictly erosional and display a sequence with increasing flow strength analogous to that of lower- and upper-regime bedforms for sand-bed channels.

In cohesive streambeds, straight longitudinal grooves and ridges form initially (Allen, 1982). These features are parallel to mean flow direction and reflect streaks in the viscous sublayer. Flute marks appear as velocity increases. Flute marks are shaped like the depression in a spoon and reflect a site where locally enhanced near-bed turbulence results in differential bed erosion and flow separation (Bridge, 2003). Further velocity increase creates transverse ridge marks shaped like ripples, which reflect supercritical flow (Bridge, 2003).

As with depositional bedforms in non-cohesive sediments, hydraulics can be inferred from the type and dimensions of erosional bedforms in cohesive sediments. Richardson and Carling (2005) presented a catalog of cohesive bedforms, sometimes known as *sculpted features*, characteristic of bedrock channels (Figure S4.21).

4.4 In-channel depositional processes

Each of the bedforms discussed in Sections 4.3.1 and 4.3.2 is also a depositional feature, particularly when the bedform stops moving. Deposition occurs when the flow or shear velocity falls below the settling velocity of a particle. This is a lower threshold than that required for entrainment (Figure 4.12). Deposition can be highly localized and limited to a zone of lower velocity and flow energy, or more widespread across and down a length of channel.

Local deposition occurs in association with isolated roughness features such as protruding boulders, instream wood, or aquatic or riparian

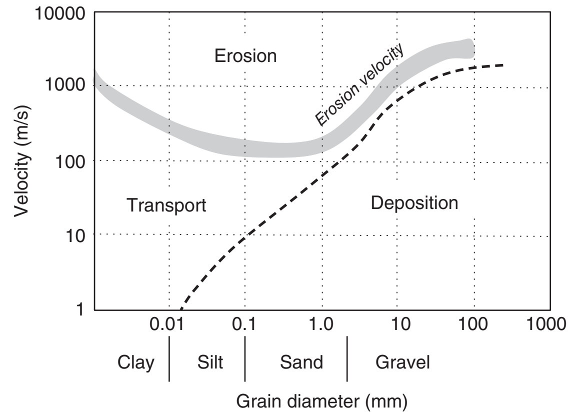

Figure 4.12 Hjulström's diagram of erosion, transport and deposition thresholds for varying grain sizes as a function of velocity. (From Hjulström, 1935.)

vegetation, or with irregularities in the channel margins such as tributary junctions or bends. Local deposition alters boundary roughness, grain-size distributions at the channel surface, and subsequent particle entrainment. Local deposition that occurs in isolated but numerous locations across the channel produces features such as particle clusters, ripples and dunes, and various types of bars.

The size and spacing of bars, in particular, can change with sediment supply and flow energy. An example comes from a headwater tributary of the River Tay in Scotland, where a sediment wave generated by lateral channel movement caused bar aggradation in some reaches. Bar aggradation resulted in flow confinement and erosion from adjacent and downstream portions of the channel, as well as trimming of bar margins, thus limiting the growth of the bars (Wathen and Hoey, 1998).

Deposition across the entire channel is facilitated by channel-spanning obstructions such as logjams and beaver dams. These obstructions can create sufficient backwater to accumulate a wedge of sediment as thick at the downstream end as the obstruction is tall, with the wedge tapering upstream in a manner dictated by channel gradient and sediment supply (Figure S4.22). Sediment accumulating upstream from the obstruction is likely to be finer grained than sediment elsewhere on the streambed, enhancing the diversity of aquatic habitats and the storage of fine organic material. The pressure gradient associated with backwater at the obstacle also enhances hyporheic exchange within the deposited sediment (Hester and Doyle, 2008). A very large obstruction such as a sediment dam associated with a rockfall, landslide, or debris flow creates a correspondingly large deposit, although natural dams caused by slope instability typically breach within a year (Costa and Schuster, 1988). Some types of large-scale natural obstructions, such as lava dams, can persist for 10^2–10^3 years, but many smaller-scale obstructions such as logjams or landslides may be present for less than a year or at most a few years.

Some portions of the channel that serve as depositional sites at lower discharge can be scoured as discharge increases. Pools typically accumulate finer sediment during low flow, but converging flow off a riffle or a central jet developing in a laterally constricted pool results in removal of sediment during high flows. Similarly, lee deposits accumulating around protruding clasts during lower discharges can be mobilized as hydraulic forces around the clasts change with increasing discharge (Thompson, 2008).

Riparian vegetation can be especially effective in facilitating deposition within the channel and along the channel margins, and increases in the density of vegetation can facilitate sediment accumulation and channel narrowing (Tal et al., 2004). An example comes from the braided, gravel-bed Tagliamento River of Italy, where Gurnell and Petts (2006) documented how deposition of large pieces of driftwood or entire trees that are capable of regrowth influences sediment deposition. These trees are deposited during waning flood stage, grow quickly and, where sufficient sand and finer sediment is available, facilitate island and floodplain development by trapping sediment and allowing more vegetation to become established.

Establishment of riparian vegetation can initiate a self-enhancing feedback as plants trap and stabilize sediments and organic matter than can provide germination sites for other plants, which in turn reinforce the development of floodplains, islands, and other landforms. Gurnell et al. (2012) referred to this process as *vegetation-mediated landform development*.

Changes in vegetation density can reflect natural fluctuations in flood magnitude and frequency. Working on ephemeral and intermittent channels in the semiarid steppe of the US Great Plains, Friedman and Lee (2002) documented repeated alternation in channel geometry and riparian vegetation through time. Channel widening and removal of vegetation occurred over a period of hours during infrequent floods caused by convective storms. Post-flood narrowing associated with regrowth of riparian forests occurred over decades between floods. The channels alternated through time between broad, braided, unvegetated rivers and narrow, sinuous, forested rivers.

In many cases, vegetation density changes as a result of invasive exotic species (Graf, 1978) or a reduction in flood peaks associated with flow regulation that allows germinating seedlings to successfully colonize lower portions of channel margins. As a result of flow regulation, larger, perennial rivers in the US Great Plains metamorphosed from braided, unvegetated rivers present in the early nineteenth century, to sinuous or anastomosing, densely forested rivers by the mid-twentieth century (Nadler and Schumm, 1981).

Cross-channel, longitudinally continuous deposition typically represents a reduction in transport capacity or an increase in sediment supply. Transport capacity can decline as a result of decreased discharge or decreased gradient. Sediment supply can increase following climate change, hillslope instability, volcanic eruptions, or changes in land cover or land use. The 1991 eruption of Mount Pinatubo, for example, covered a third of the Pasig-Potrero River watershed in the Philippines with sand-sized pyroclastic flow deposits. This abrupt change in sediment supply caused the gravel-bedded river to become a planar, highly mobile, sand-bedded channel that gradually transitioned back toward a gravel channel over the succeeding 15 years (Gran et al., 2006).

Cross-channel, longitudinally continuous deposition can occur as a relatively minor or temporary effect that results in finer sediment being deposited on the surface of a coarse-grained channel

(Figure S4.23). Although the effects on cross-sectional geometry can be minor, fining of the bed sediment can alter hyporheic exchange and habitat suitability for benthic organisms such as algae, aquatic insects, or larval fish.

More substantial and sustained deposition throughout a channel is known as *aggradation* and can result in sufficient loss of channel cross-sectional area to cause enhanced overbank flooding and lateral channel movement or the formation of secondary channels. A common cause of aggradation leading to loss of cross-sectional area is the clearance of native vegetation and start of agriculture within a drainage basin, as discussed in more detail in Supplemental Section 4.4.

4.5 Bank stability and erosion

The ability of flow to erode the banks of a channel, and the processes and rates of bank erosion, strongly influence sediment supply, channel geometry, riparian vegetation, and rates of lateral channel migration. Bank erosion thus represents, like bedforms, a means of adjusting channel geometry, external resistance to flow, and sediment supply.

Bank erosion can result in property loss and damage to infrastructure such as buildings and bridges. Human alterations of water and sediment supply to the channel commonly result in altered rates of bank erosion. Even unaltered, "natural" bank erosion is in many cases regarded as an indicator of channel instability and a sign that mitigation is needed to limit bank erosion. This represents a fundamental misunderstanding of river processes. A stable channel transporting bed material must entrain an amount of sediment equivalent to that deposited. Most in-channel deposition occurs in bars, and most "compensating" erosion comes from banks as the channel expands to maintain conveyance. Consequently, local lateral instability in the form of local bank erosion is a natural feature of stable rivers transporting bed material, but is commonly misinterpreted as a sign of general channel instability. Given the tendency to build immediately adjacent to channels and then attempt to prevent bank erosion, much effort is devoted to analyzing and modeling bank stability.

As with bed erosion, the forces driving sediment removal from banks must exceed the resisting forces for banks to erode. This becomes complicated to assess in part because stream banks are commonly layered, with coarser sediment at the bottom as a result of deposition within the channel and finer sediment toward the top of the bank as a result of overbank deposition (Figure S4.24). Bank sediment typically grows progressively finer grained downstream or along channel segments of lower gradient, but banks can be very spatially heterogeneous. The finer sediments in the upper bank add cohesion, and the roots of riparian vegetation can further enhance the resistance of the bank sediment to erosion (Figure S4.25).

The strength of stream banks reflects the frictional properties of the bank sediment, effective normal stress, and effective cohesion (Simon et al., 1999). *Effective normal stress*, σ', is the difference between normal stress (the perpendicular component of total stress, σ) and pore pressure, u: in other words, $\sigma' = \sigma - u$. Effective normal stress results from static friction that keeps particles together in a stream bank minus the effect of pore pressure that keeps particles separate. *Effective cohesion* results from true cohesive forces, from matric suction within the unsaturated portion of the bank, and/or from root reinforcement via vegetation (Eaton, 2006). Acting against these resisting forces are the driving forces of gravity and hydraulic force expressed as shear stress.

Pore water content and pressure is one of the more important influences on bank stability (Rinaldi and Darby, 2008), and involves multiple effects. Pore water reduces shear strength by increasing lubrication between sediment particles. Pore water increases the unit weight of the bank material, making unsupported material more susceptible to failure. Pore water destabilizes banks by facilitating the presence of water in tension cracks. Pore water also creates seepage forces that can stabilize or destabilize the banks.

Pore water pressure results from the pressure of water filling the voids between particles. Negative pore water pressures above the water table reflect the surface tension of pore water in voids, which creates a suction effect on surrounding particles and stabilizes the banks. Positive pore water pressures below the water table, in contrast, help to force particles apart and destabilize banks. Pore water pressures are extremely transient in response to changes in precipitation and stream flow, but they typically promote bank failure during the falling limb of the hydrograph when the bank sediment is at or near saturation and the confining pressure of the river water is removed.

Vegetation also influences stream banks in diverse ways (Merritt, 2013). Plants increase the mass of stream banks and can thus facilitate bank failure. However, plants near the water's edge create flow resistance that reduces near-bank velocity and shear stress and increases bank stability (Griffin et al., 2005; Gorrick and Rodríguez, 2012). Plant roots increase the resistance of the sediment to shearing (Pollen and Simon, 2005). And, plants alter bank pore water pressure by affecting infiltration, evaporation, and transpiration (Griffin et al., 2005; Pollen-Bankhead and Simon, 2010). Vegetation exerts the greatest influence on bank stability along low, shallow banks in weakly cohesive sediments and along small channels (Eaton and Millar, 2004).

Not all vegetation is created equal with regard to bank stability. Channels with forested banks tend to be wider than channels in grasslands for at least two reasons. First, wood recruited to the channel from the riparian zone promotes channel widening. Second, grass grows readily on point bars, facilitating more rapid deposition of suspended sediment and narrower channels (Allmendinger et al., 2005). Among woody vegetation, densely growing species such as willows (*Salix* spp.) can limit bank erosion and channel widening more effectively than species in which individual plants are more widely dispersed (David et al., 2009) (Figure S4.26).

Cyclic bank erosion can occur over decades when small volumes of sediment are removed between large trees on the bank, creating a scalloped bank morphology buttressed by large trees. The trees are gradually undercut and topple into the channel, resulting in a larger volume of bank erosion and a new round of the cycle (Pizzuto et al., 2010) (Figure 4.13).

Figure 4.13 Scalloped bank along a small creek in Connecticut, USA, caused by locally increased bank resistance as a result of tree roots.

The primary processes eroding stream banks involve hydraulic action or mass failure. Fluvial detachment via hydraulic action is likely to be more important in coarse-grained or smaller rivers. Mass failure becomes progressively more important downstream as bank heights increase and bank sediments become finer grained (Lawler, 1992).

Hydraulic action results from shear stress exerted against a bank and is related to the near-bank velocity. Individual particles can be detached from the bank face in non-cohesive or weakly cohesive sediment and the base of the bank can be preferentially eroded (Rinaldi and Darby, 2008). Even cohesive sediments can be made more susceptible to hydraulic action by processes that weaken and detach sediment, such as shrink–swell or freeze–thaw cycles (Wynn et al., 2008). Conversely, flow resistance associated with riparian vegetation can reduce bank erosion through hydraulic action. Hydraulic action is typically quantified using an excess shear stress equation similar to Equation 4.14, with fluvial bank erosion rate per unit time substituted for bed load transport rate (Rinaldi and Darby, 2008).

Processes that weaken and detach bank sediment also promote mass failure through slumping or toppling of a slab. Seepage also reduces the cohesion of bank sediment by removing clay and, in some cases, promoting piping in the bank. Seepage can be particularly effective where rapid reductions in flow stage leave saturated banks without lateral support, a situation common in rivers regulated for hydropower generation. Trampling by large numbers of grazing animals congregating along streams can further enhance bank erosion by breaking down the banks and creating hydraulic roughness (Trimble and Mendel, 1995).

Mass failure of banks takes several forms (Osman and Thorne, 1988; Knighton, 1998). Shallow slips in which relatively thin segments of bank detach and then disintegrate dominate in non-cohesive sediment. Slab-type failures in which a vertical slab detaches from the bank and then topples are most common in weakly cohesive sediment and near-vertical banks. Deep-seated rotational slips dominate in cohesive sediment. In these failures, a mass of sediment slides down the bank along a curved failure surface, rotating so that the toe of the failure protrudes into the river.

All forms of mass failure are enhanced by scour at the base of the bank that oversteepens the bank. In strongly layered banks, removal of the coarser, non-cohesive sediment in the lower bank can create overhangs that eventually collapse as blocks into the channel. Such blocks can either quickly break up and disperse, or persist for more than a year, in part because of the dense, fine plant roots within the block. Persistent blocks act similarly to boulders and can protect the lower bank from further erosion or deflect the current toward the bank toe and enhance basal scouring (Figures 4.14 and S4.27). Bank collapse via overhanging blocks is particularly common in permafrost regions where slow thawing of the soil ice facilitates the persistence of collapsed blocks (Walker et al., 1987), and in wet meadows where bank erosion creates peat blocks (Warburton and Evans, 2011). Root reinforcement of bank sediment by riparian vegetation can limit mass failure, particularly where the vegetation grows along the bank toe or at the intersection of the failure plane with the floodplain surface (Van De Wiel and Darby, 2007).

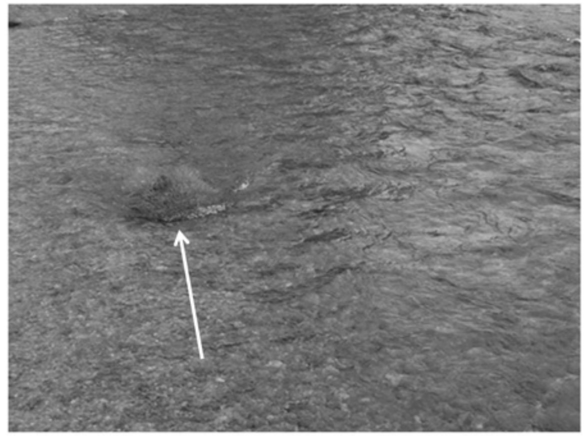

Figure 4.14 Fine-grained bank collapse block that will act as an isolated roughness element until the vegetation holding the sediment together decays. This block, indicated by the white arrow parallel to flow, is about 40 cm in diameter, and is much larger than the average size of the bed substrate.

Cold-region rivers with a seasonal ice cover can have a distinctive bank morphology in the form of a two-level bank structure that reflects ice scouring (Boucher et al., 2012). Elevation of the ice surface during freeze-over and jamming of ice blocks during break-up lead to abrasion of the banks by ice and the formation of a steep segment of bank above the bankfull stage. Ice gouging and overbank sedimentation during ice-jam flooding can create an elevated ridge or bench along some rivers, which is referred to as a *bechevnik* from the Russian word for tow rope, because these benches formed convenient paths for towing boats upstream along Siberian rivers (Ettema and Kempema, 2012).

Several models quantify and predict bank stability, as reviewed by Pizzuto (2003) and Rinaldi and Darby (2008). Models can be differentiated as *mechanistic models* based on the physics of particular erosional processes, and *parametric models* that relate bank erosion to potential controlling parameters—typically near-bank velocity or shear stress—using empirical coefficients (Pizzuto, 2003).

Bank erodibility parameters are modeled using methods similar to those for entrainment of bed sediments, with modifications to account for the effect of bank angle on the downslope component of particle weight and for partly packed and cemented sediments (Rinaldi and Darby, 2008). Near-bank hydraulic force is either directly measured or estimated using hydraulic models. Models developed for non-cohesive banks include the mechanistic model of Kovacs and Parker (1994), which uses a bed load transport model to compute bed load transport on steeply sloping banks coupled with the near-bank velocity. Because vegetation, moisture, and even small amounts of fine sediment add cohesion to banks, most models focus on cohesive banks.

Mechanistic models for cohesive banks are generally 2D models that evaluate bank stability in terms of the soil strength and bank geometry (Pizzuto, 2003). For example, Simon et al. (2000) developed a bank stability algorithm and the mechanistic bank stability and toe erosion (BSTEM) model for layered, cohesive stream banks. They combined the Coulomb equation for saturated banks with the Fredlund et al. (1978) equation for unsaturated banks. The Coulomb equation is

$$S_t = c + \sigma' \tan\phi \tag{4.18}$$

where S_t is shear strength, c is cohesion, ϕ is the angle of internal friction, and σ' is the effective normal stress. Total normal stress, which tends to hold sediment together, is the sum of effective normal stress and pore pressure. Under saturated conditions, pore pressure is positive and effective normal stress is lower. In partially saturated soils, effective normal stress is increased. Consequently, bank sediment is more susceptible to mass failure under saturated conditions (Robert, 2003).

The failure criterion of Fredlund et al. (1978) is

$$\tau = c' + (\sigma - u_a)\tan\phi' + (u_a - u_w)\tan\phi^b \tag{4.19}$$

where τ is the shear strength, c' is the effective cohesion, σ is the normal stress, u_a is the pore air pressure, ϕ' is the friction angle in terms of effective stress, u_w is the pore water pressure, $(u_a - u_w)$ is the matric suction, and ϕ^b is the angle expressing the rate of increase in strength relative to the matric suction.

Parameter uncertainties in bank stability models are typically so large as a result of natural variability in the parameters that the likelihood of generating unreliable predictions exceeds 80% (Samadi et al., 2009). In addition, processes that weaken and strengthen banks interact in sometimes unpredictable manners. Consequently, rather than use a deterministic model with a single value for each bank material property, a probabilistic representation of effective bank material strength parameters may be the most appropriate approach (Parker et al., 2008).

Supplemental Section 4.5 discusses measuring bank stability and erosion.

4.6 Sediment budgets

A sediment budget quantifies fluxes of sediment past a given point in a watershed. Sediment budgets are based on the very simple formula of inputs (I) minus the change in mass or volume of sediment (ΔS) stored in the channel reach during a specified time interval,

$$I - \Delta S = \varphi \tag{4.20}$$

where φ is the mass or volume of sediment output from the channel reach during the specified time interval. φ is also known as *sediment yield*. Sediment inputs represent those coming from upstream and subsurface sources on the main channel and tributaries, as well as from sources beyond the channel, including hillslopes, glaciers, terraces and other valley-bottom deposits, and eolian inputs. Storage occurs within the channel bed, banks, bedforms, and in overbank areas such as the floodplain. Output is transported within the channel as dissolved, wash, suspended, and bed load (Figure 4.15).

The mathematical simplicity of Equation 4.21 is deceptive because each of the three primary variables can fluctuate greatly through time, across

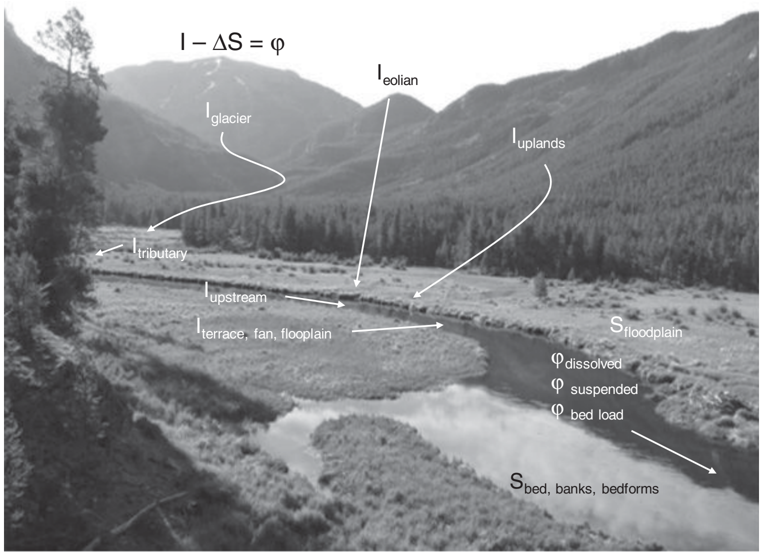

Figure 4.15 Schematic illustration of the basic components of a sediment budget. Inputs *I* to a river segment can come from uplands and glaciers, eolian deposition, mainstem and tributary transport, or erosion of terraces, alluvial fans, floodplains, and other valley-bottom deposits. Storage *S* of sediment can occur in the floodplain or in various portions of the channel. Sediment outputs φ occur via downstream transport as dissolved, suspended, and bed load.

a drainage basin and between basins, making it extremely difficult to accurately quantify input, storage and output even over relatively short time intervals. Sediment inputs can be gradual, as in slope wash, average tributary sediment transport, or soil creep. Or, sediment inputs can be abrupt, as in debris flows or tributary flash floods. Inputs can also be seasonally driven, aperiodic, or variable over time spans of thousands of years because of fluctuations in base level, climate, or land use.

Sediment inputs tend to be temporally and spatially heterogeneous in a wide variety of basins. Research in mountainous basins provides examples: 75% of the long-term sediment flux in Taiwan occurs during typhoon-generated floods occupying <1% of the flow duration curve (Kao and Milliman, 2008). Over a period of 70 years, half of the sediment load from a river in California, USA, was delivered in less than 5 weeks (Farnsworth and Milliman, 2003). Sediment yield in drainages with periodic disturbances such as volcanic eruptions and wildfires is likely to be dominated by episodic inputs from both disturbed hillslopes and continuing channel instability (Gran et al., 2011).

In large river basins as diverse as the Amazon and the Mississippi, one portion of the catchment supplies the great majority of total sediment output (Figure 4.5), as noted earlier. In the case of the Amazon, the Andes of Peru and Bolivia supply more than 80% of the sediment load but constitute only ~10% of the basin area (Meade et al., 1985; Meade, 2007).

The relative importance of diverse sediment sources varies greatly among basins. High-relief basins tend to be dominated by hillslope sources. Laterally mobile rivers with extensive floodplains can obtain sediment predominantly from bank erosion. High-latitude basins can be dominated by glacial sources (Gurnell, 1995).

The relative importance of diverse sediment sources also varies among solute, wash, suspended, and bed load. In most catchments, solutes come primarily from groundwater inputs. Bank sediments can contribute nearly 40% of the total suspended sediment, even in relatively low-energy catchments (Walling et al., 1999), with up to 80% of the total suspended sediment yield coming from bank sources in some highly unstable, incised channel networks (Simon and Darby, 2002). Bed load can come primarily from mobilization of streambed sediments, inputs from smaller, steeper tributary catchments or, particularly in high-relief catchments, from upland sources such as hillslope mass movements.

The predominant source of sediment can also vary through time. This is illustrated by the Upper Mississippi River, USA, where the dominant source of suspended sediment shifted from agricultural soil erosion in the mid-twentieth century to accelerated erosion of stream banks since 1980 as discharge has increased from climate change and channelization (Belmont et al., 2011).

The existence of sediment storage is reflected in the *sediment delivery ratio*. This ratio represents the difference between volume of sediment generated and volume of sediment stored or transported from the basin, typically as a function of drainage area, A

$$\gamma = \alpha A \varphi \tag{4.21}$$

where γ is sediment delivery ratio and α and φ are empirical parameters. Sediment delivery ratio is typically calculated from observed sediment yields (y) and measured or estimated gross erosion rates (e):

$$y = \gamma e \tag{4.22}$$

Combining Equations 4.22 and 4.23 results in

$$y = bA^{\theta} \tag{4.23}$$

where y is areal average sediment yield and θ is an empirical parameter that varies from -0.52 to 0.12 (Milliman and Meade, 1983; Lu et al., 2005).

Sediment delivery ratio is typically between 0 and 1, indicating that only a fraction of the total sediment detached from eroding sources actually leaves a catchment. Although sediment delivery ratio is commonly related to drainage area, other factors such as topography, land cover, land use, lithology, and cyclic channel processes such as gully erosion and filling can strongly influence sediment delivery ratio (Wohl, 2010). In general, smaller, steeper catchments have a greater sediment delivery ratio because such catchments have proportionally less sediment storage in features such as floodplains.

All rivers with large sediment loads originate in mountains, and the majority of sediment load within most river basins comes from the mountainous portion of the basin (Milliman and Syvitski, 1992). The correlation between sediment load and small, steep catchments reflects the fact that such catchments typically have higher within-hillslope connectivity, hillslope–channel connectivity, and within-channel connectivity for surface water and sediment. Steep hillslopes, numerous first-order channels and steep, narrow valley bottoms limit sediment storage in colluvial (hillslope) and alluvial sites, creating efficient delivery of sediment into and through river networks. Magnitudes and timescales of sediment storage increase in larger basins where wide valley bottoms buffer channels from hillslope inputs and provide larger storage areas with terraces, floodplains, and alluvial fans.

The influence of climate on sediment inputs appears in plots of sediment yield versus a variable such as effective precipitation. (Effective precipitation is the precipitation that actually contributes to runoff.) Such plots originated with Langbein and Schumm (1958), who used sedimentation records from the United States to demonstrate that sediment yield peaks at ~300 mm effective precipitation. Lesser values of precipitation produce insufficient runoff to mobilize large quantities of sediment. Continuous vegetation cover effectively limits erosion at higher precipitation values. Subsequent work indicates a secondary sedimentation peak in the seasonal tropics. Despite the mostly continuous vegetation cover in the tropics, the great intensity of precipitation may limit the ability of vegetation to retard runoff and surface erosion or, in steep terrains, mass movements such as landslides.

The influence of lithology on sediment yield reflects relative rates of bedrock weathering and erosion, as well as the characteristics of weathering products. Other factors being equal, more erodible and/or clastic sedimentary rocks tend to yield greater sediment volumes than rocks more resistant to weathering and erosion. Chemical sedimentary rocks such as limestone also typically have low yields of granular sediment because most weathering products move as dissolved load.

The dominant influence on sediment yield in many catchments is land use. Hooke (2000) estimated that humans move more sediment globally than any other geomorphic agent, and sediment yields from areas where natural vegetation and topography have been altered are among the highest in the world. Supplemental Section 4.6 discusses in more detail how diverse human activities influence sediment yield and how sediment yield is measured or estimated.

The volume and residence time of sediment storage in depositional features can also vary greatly. Residence time of floodplain sediments, for example, reflects the balance between rates of sediment accumulation through vertical and lateral accretion, rates of floodplain erosion through bank migration, and the size of the floodplain (Section 6.1). The mainstem Amazon in Brazil is bordered by ~90,000 km^2 of floodplain, and floodplain width increases downstream. Mertes et al. (1996) estimated mean residence times for floodplain sediments in this portion of the river as being ~1000–2000 years, and reaching periods as long as 10,000 years.

Residence time of sediment in floodplains and along river networks typically increases with drainage area. At larger drainage areas, flows of sufficient magnitude to "turn-over" (i.e., erode) the entire floodplain and the majority of storage sites along the channel are very rare (Harvey, 2002), in part because precipitation capable of creating a high percentage of contributing area within a drainage becomes infrequent at very large drainage areas. A relatively small convective cell can produce heavy rainfall over all of a small drainage basin, increasing the chances that the entire basin will be contributing and that widespread erosion will occur. Such basin-wide precipitation is more common in small basins, creating shorter residence times for sediment and higher delivery ratios.

Sediment output varies not only with the occurrence of precipitation and flows capable of mobilizing sediment in colluvial and alluvial storage areas, but also with the duration of flows capable of transporting the sediment. Once the entrainment threshold in a channel is exceeded, sediment transport tends to increase with flow duration. The magnitude of the entrainment threshold thus exerts an important control on sediment output. The majority of the sediment output in sand-bed channels with relatively low entrainment thresholds can occur during relatively frequent flows (Wolman and Miller, 1960). In contrast, the majority of sediment output in very coarse-grained or otherwise resistant-boundary channels may occur during infrequent, high magnitude flows (Wohl, 1992).

Case studies illustrate some of the spatial and temporal complexities of sediment budgets.

- Trimble (1983) examined historical changes in the sediment budget of Coon Creek in Wisconsin, USA, using: measured suspended and bed load yields from Coon Creek, reservoir sedimentation data from a nearby analogous basin, estimates of upland sheet and rill erosion based on the Universal Soil Loss Equation, sediment cores from valley bottoms, and mapped historical accumulations of sediment from aerial photographs. Examining the period 1853–1975, he estimated that about half of the upland sediment brought into the river network as a result of agricultural development went into floodplain storage, with less than 7% of the sediment transported out of the basin.
- Similarly, Fryirs and Brierley (2001) found that lowland portions of the 450 km^2 Bega River catchment in southeastern Australia have stored more than 80% of the sediment mobilized from the uplands since European settlement of the catchment in the mid-nineteenth century. Fryirs and Brierley used historical ground and aerial photographs, stratigraphy and sedimentology of valley bottoms assessed from geomorphic mapping and sediment cores, and sediment delivery to the estuary to develop an alluvial sediment budget through time for sand and fine gravel in the catchment.

These examples illustrate how strongly depositional features can influence sediment yields and sediment budgets. In this context, depositional features such as floodplains reduce upland-channel and river network sediment connectivity by storing substantial amounts of sediment.

Supplemental Section 4.6 discusses methods of measuring and modeling sediment budgets and outputs.

4.7 Summary

Natural channels differ from engineered channels in being able to adjust boundary roughness, channel geometry, and sediment transport in response to changes in water and sediment supplied to the channel. Most of these adjustments involve sediment dynamics—the entrainment, transport, and deposition of individual particles and aggregates of particles.

Historically, quantification of sediment dynamics began with the simple scenario of a channel bed with relatively uniform, sand-sized grains that could be readily collected and sieved to quantify the grain-size distribution. Sand-bed channels have a relatively low entrainment threshold and can be adequately represented by bulk sediment transport equations. The entrainment and transport equations developed for sand-bed channels are based on estimation of the magnitude of shear stress above a critical value necessary to mobilize sediment, as formulated by Shields. These equations remain the starting point for characterizing sediment dynamics.

Quantitative studies of sediment dynamics have expanded to include a progressively broader range of natural channels, from fine-grained cohesive beds, to bedrock beds, and gravel or boulder beds. Entrainment and transport in these types of channels are not adequately approximated with an excess shear stress relation for entrainment or a bulk transport relation because of effects such as shielding and protrusion and limited sediment supply. Mobilization of particles from beds of mixed grain sizes occurs in phases. Equations using two-fraction transport or grain–grain interactions better capture the processes operating in these channels.

The distinction between readily mobile and infrequently mobile bedforms reflects basic differences in thresholds of entrainment and mobility between sand-bed and other channel types. The readily mobile bedforms of ripples, dunes, and antidunes respond to a range of flow magnitude. Infrequently mobile bedforms, including particle clusters, transverse ribs, step–pool and pool–riffle sequences, change most actively under relatively high flow magnitudes.

Sediment can also be eroded from or deposited along stream banks. As with streambeds, bank erosion reflects the frictional properties of the sediment. Bank erosion is also strongly influenced by bank stratigraphy, pore water pressure, and riparian vegetation, and is more likely than bed erosion to take the form of mass failure. As with bed sediment, efforts to quantitatively model and predict bank erosion are limited by large spatial and temporal variability in sediment properties and in the forces acting on the sediment.

A mass balance of sediment in the form of a sediment budget reflects the inputs, storage, and outputs of sediment across a reference area that can vary from a small subcatchment to the Amazon. Regardless of spatial scale or geographic setting, sediment budgets reveal that sediment comes disproportionately from limited areas of a catchment, and moves disproportionately during limited intervals of time. Sediment export includes solute, wash, suspended, and bed load, but the majority of sediment moved from most river catchments is transported in suspension.

Chapter 5
Channel forms

Channel form can be examined at many levels, as reflected in diverse classifications for rivers. The predominant bedforms present in a river can be used to distinguish step–pool from pool–riffle channels (Section 4.3.2), for example, or the focus on channel form can be at the cross-sectional scale, based on descriptors such as width/depth ratio or exponents of at-a-station hydraulic geometry relations, both of which are introduced in this chapter. Channel planform is commonly used to distinguish straight, meandering, braided, and anabranching channels with differing degrees of sinuosity and single- or multi-thread channels. Channel form can also include the vertical dimension over the entire length of a river, as when longitudinal profiles are categorized in terms of concavity or stream gradient indices. Just as river classifications attempt to identify consistent thresholds that distinguish channel forms existing within a continuum, the distinction of cross-sectional, planform, and longitudinal channel forms in this chapter represents an arbitrary division of aspects of river morphology that intergrade with and influence one another.

5.1 Cross-sectional geometry

Process and form in natural channels typically exhibit large spatial and temporal variability. Much work has been devoted to identifying parameters that can usefully describe the mean state and the variations in process and form. In the preceding chapter, channel form was characterized in terms of the dominant bed material (e.g., sand-bed channel, gravel-bed channels) or the predominant bedforms (e.g., pool–riffle channel, step–pool channel). Channel form can also be described with respect to cross-sectional geometry, with a focus on mean cross-sectional characteristics.

Cross-sectional form parameters such as width, depth, cross-sectional area, wetted perimeter, and hydraulic radius (Figure 5.1) are typically measured at bankfull stage, under the assumption that bankfull stage defines a channel area that is filled by flow at least once every 2 years, or more frequently. Bankfull flow is a problematic concept that is used quite differently in different studies, but nonetheless forms the standard for measuring cross-sectional geometry. Cross-sectional parameters can also be measured for specific flow magnitudes, such as the mean annual flood or base flow.

5.1.1 Bankfull, dominant, and effective discharge

Bankfull discharge is one of the most widely used reference discharges in fluvial geomorphology, yet the definition and implications of this flow remain

Rivers in the Landscape: Science and Management, First Edition. Ellen Wohl.

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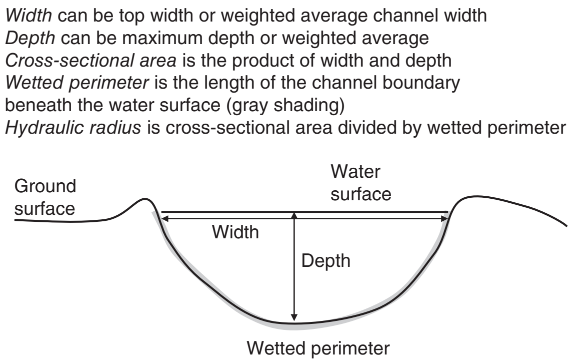


Figure 5.1 Downstream view in a channel, indicating cross-sectional geometry form parameters.

controversial. Wolman and Leopold (1957) defined bankfull discharge as the stage just before flow begins to overtop the banks. Bankfull discharge has subsequently assumed various implications because it has been equated to a specific recurrence interval and geomorphic function. Numerous studies indicate that a discharge that nearly overtops the banks recurs approximately every 1–2 years on many channels (Leopold et al., 1964; Castro and Jackson, 2001). Consequently, bankfull discharge is sometimes defined based on recurrence interval rather than stage with respect to channel morphology. Because a discharge that recurs every 1–2 years transports the majority of suspended sediment in many rivers (Simon et al., 2004), bankfull discharge has been interpreted as the most important flow magnitude for controlling channel process and form (Wolman and Miller, 1960; Dunne and Leopold, 1978).

Problems arise in that bankfull discharge can be difficult to define based on channel morphology. Designating bankfull stage is relatively straightforward where a clearly defined, regular top of bank separates the channel from overbank areas. Bankfull stage can be difficult to define morphologically if a river has inset channels. The channel may be so deeply incised, for example, that the top of the bank has little relevance to flow volume. The top of the bank may be at different levels on each side of the channel. The channel can also display multiple convexities along its side slopes that reflect different flow magnitudes. Comparisons of morphological definitions using various criteria proposed in individual studies—top of bank, bank inflection, ratio of channel width to mean depth, level of significant change in the relation between wetted area and top channel width, and first maximum local bank slope—indicate that, on average, discharge estimates vary by a factor of three at a given site (Radecki-Pawlik, 2002; Navratil et al., 2006).

Another source of uncertainty is that bankfull discharge defined in relation to a channel morphologic feature can have very different recurrence intervals among different sites. In hydroclimatic regions with extreme annual and interannual flow variability, such as arid and semiarid regions, the morphologically defined bankfull flow may have a recurrence interval much longer than 1–2 years. Recurrence intervals associated with a consistent morphological feature can vary by a factor of two between channel segments within regions as diverse as Puerto Rico (Pike and Scatena, 2010) and snowmelt rivers in the US Rocky Mountains (Segura and Pitlick, 2010).

Finally, whether defined from channel morphology or recurrence interval, bankfull discharge does not necessarily transport the majority of suspended sediment or exert a greater influence on channel form than other magnitudes of flow (Wohl, 2010). Analyses of suspended and bedload transport on a variety of rivers indicate that discharge with a recurrence interval of 1–2 years transports much of the sediment moved in many rivers, but there are exceptions to this generalization, and channel morphologic features are likely to reflect multiple recurrent discharge magnitudes (Pickup and Warner, 1976; O'Connor et al., 1986; Turowski and Rickenmann, 2009).

As a broad generalization, large magnitude, infrequent flows are more likely to strongly influence sediment transport and channel morphology in rivers with higher thresholds for mobilizing bed and bank materials, and greater temporal variation in discharge magnitude. In other words, sediment transport and channel morphology are more likely to reflect larger, infrequent flows in gravel-bed, boulder-bed, and bedrock rivers, and rivers in very dry or seasonal tropical climatic regimes.

The concept of a *dominant discharge*, which is sometimes equated with bankfull discharge, reflects the idea of a single flow magnitude which, if maintained, will maintain the same average dimensions and channel morphology as those that result from a stable stream's entire hydrologic regime (Crowder and Knapp, 2005). When used in numerical modeling, dominant discharge is the flow that moves the same amount of bed material, with the same size distribution, as the actual flow regime. The assumption is that channel morphology under a dominant discharge will also remain similar to the morphology under the actual flow regime.

Effective discharge is the discharge that transports the largest amount of sediment (Schmidt and Morche, 2006). This magnitude of discharge may or may not equate to bankfull stage.

Bankfull discharge, effective discharge, mean annual flood, and the 1.5-year flow have all been proposed as constituting dominant discharge (Rosgen and Silvey, 1996; Griffiths and Carson, 2000). The idea that a single magnitude of flow strongly dominates river process and form is an oversimplification of the complexity of interactions among flow, sediment, and channel geometry through time. This oversimplification can have important consequences when used in river restoration as an index value for scaling channel dimensions and designing stable channels, particularly if excessive emphasis on dominant discharge results in neglecting the geomorphic and ecological importance of less frequent, larger flows or of base flows.

5.1.2 Width to depth ratio

The ratio of channel top width to mean flow depth (*w/d*) is one of the most commonly used parameters of cross-sectional geometry. The *w/d* ratio can be used to infer the limit strength and relative erodibility of bed and bank materials, relative base level stability, and consistency of water and sediment inputs.

Channels of diverse size flowing on similar bed materials become wider more rapidly than they become deeper as discharge increases with time at a cross section or proceeding downstream in a river network. Widening a channel requires less sediment transport capacity than deepening a channel. Bank erosion simply requires eroding the banks, but the resulting sediment can then be stored in the channel. Bed erosion requires both entraining the bed sediment and transporting the sediment downstream. Banks are also more likely to become unstable as they grow taller, unless they are in very cohesive material, so that deepening a channel typically results in associated widening as over-steepened banks collapse. In addition, the channel bed typically consists of coarser sediment than the banks and thus requires more flow energy to entrain and remove.

Relative erodibility of bed and bank materials can reflect differences in properties such as cohesion or grain size between the bed and banks that influence the ability of stream flow to remove boundary material. As banks become more erodible, *w/d* increases: braided channels typically have greater *w/d* ratios than single-thread channels. Low bank erodibility can reflect high percentages of bedrock or cohesive sediment in the bank, or effective bank stabilization by vegetation.

If other factors are equal, channels with forested banks tend to be wider and have lower rates of bank erosion and channel migration than channels with grassy banks (Allmendinger et al., 2005). As noted in Section 4.5, the degree to which forest or other vegetation influences bank stability and *w/d* ratio varies. Riparian trees with dense root networks and greater stem density that increases overbank roughness tend to result in narrower channels than channels in more open woodlands. This is illustrated by plots of bankfull width versus discharge for gravel-bed channels in Colorado, USA, (Andrews, 1984) and the United Kingdom (1986). In both cases, rivers with banks heavily vegetated with trees and thick brush had narrower channels for a given unit discharge than did rivers with banks primarily covered in grasses and brush (Figure S5.1).

The effect of dense riparian vegetation was also demonstrated in a comparison of channel responses to increased peak flows associated with snowmaking at commercial ski resorts in Colorado, USA. Channels lined with dense willow communities showed much less erosion and widening following increased

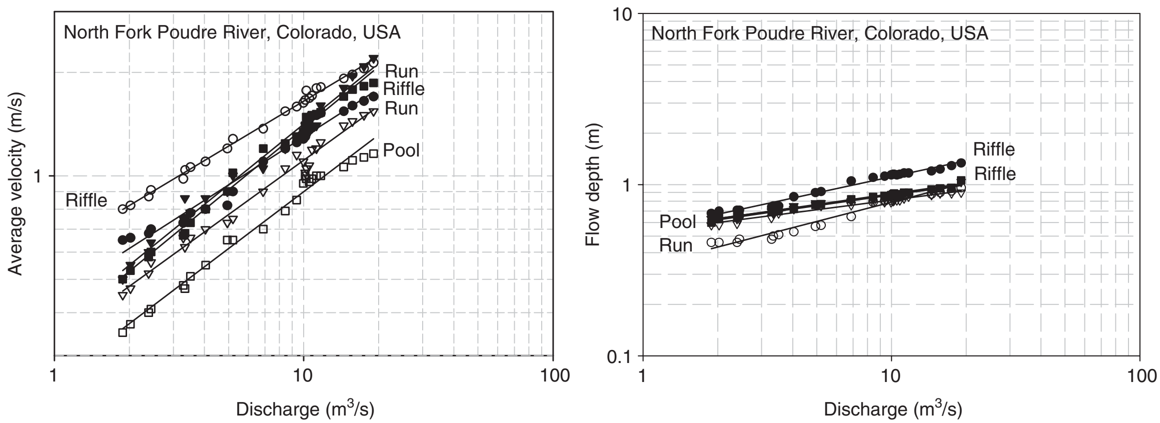


Figure 5.2 At-a-station hydraulic geometry curves for depth and velocity on the North Fork Poudre River in Colorado, USA (drainage area 980 km^2). Individual curves within each plot are for different channel units along this pool–riffle channel.

flows than channels lined with open conifer forests (David et al., 2009) (Figure S4.21).

Grazing animals can also increase channel width and *w/d* ratios indirectly by selectively grazing on and removing riparian vegetation, and directly by trampling stream banks and enhancing bank erosion. Such effects have been documented for domesticated animals (Trimble and Mendel, 1995) and for wild animals such as bison (Butler, 2006).

Channel *w/d* ratio can also reflect base level constraints. Rivers do not incise below base level, so an increase in discharge when base level remains constant is more likely to result in channel widening or some planform change such as formation of secondary channels, than in deepening of the river. Conversely, channels develop lower *w/d* ratios (i.e., deepen faster than they widen) under relative base level fall.

Channel cross-sectional geometry can also reflect changes or relative consistency in water and sediment inputs. An increase in sediment yield is likely to cause bed aggradation and channel widening, leading to a larger *w/d* ratio. A decrease in sediment yield can cause bed erosion, but is also likely to result in bank erosion, leading to less predictable changes in *w/d* ratio. Four conceptual models commonly used to describe and predict changes in cross-sectional geometry are discussed in the next four sections.

5.1.3 Hydraulic geometry

At-a-station hydraulic geometry

At-a-station hydraulic geometry is used to describe changes in cross-sectional parameters at a site with changes in flow, and to compare rates of change in these parameters between sites. At-a-station hydraulic geometry characterizes how changing discharge alters width, depth, and velocity at a cross section (Figure 5.2). Starting with the continuity equation, and assuming that discharge is the primary influence on hydraulic variables, Leopold and Maddock (1953) proposed that

$$w = aQ^b \tag{5.1}$$

$$d = cQ^f \tag{5.2}$$

$$v = kQ^m \tag{5.3}$$

where w is width, d is mean flow depth, v is mean flow velocity, Q is discharge, and a, c, k, b, f, and m are numerical constants. Based on the continuity equation, the product of a, c, and k is one, and the sum of b, f, and m is one.

The rates of change of w, d, and v reflect

- the shape and relative erodibility of the channel—channels with nearly vertical, cohesive banks, for example, have a very low rate of change in width (Knighton, 1974);

- sediment transport—channels carrying large amounts of bedload are typically wide and shallow, and depth increases little with discharge, whereas velocity increases rapidly (Wilcock, 1971);
- the slope of the water surface; and
- the roughness of the wetted perimeter—velocity increases rapidly in a channel in which increasing flow depth quickly makes the protrusion height of roughness elements a small proportion of flow depth (Ferguson, 1986; Eaton, 2013).

Studies of steep, coarse-grained channels suggest that channel segments in which velocity increases more rapidly than w and d are dominated by grain resistance, because relative grain submergence increases quickly as discharge increases, and associated flow resistance declines. Channel segments in which width and depth increase more rapidly are dominated by form resistance (David et al., 2010). Rates of change in depth and velocity in relation to resistance may not be linear, however, if different sources of resistance occur with changing stage. Although velocity may increase more rapidly once grain resistance becomes negligible, for example, form resistance associated with a mobile bed can become more important at higher flows and slow the rate of velocity increase at the largest discharges.

Although the complexity of at-a-station hydraulic geometry relations limits generalizations, the width exponent primarily reflects channel geometry and boundary composition. The depth and velocity exponents reflect cross-sectional form as well as hydraulic resistance and sediment transport, which tend to be more variable than form parameters (Knighton, 1998).

Average values of at-a-station hydraulic geometry exponents are $b = 0.23$, $f = 0.42$, $m = 0.35$ (Park, 1977). These values indicate that depth typically increases more rapidly with discharge than does width or velocity, although the exponents vary widely among channels and among different cross sections on a single channel (Reid et al., 2010). Inflection points that mark variations in the rate of increase in w, d, or v with increasing Q can reflect increasing submergence of grain or form resistance elements or the initiation of overbank flow. Hydraulic geometry curves can be divided into three portions (Knighton, 1998):

- one portion for low discharges below the threshold of sediment movement, when flow characteristics reflect a cross-sectional geometry left from earlier high flows;
- a middle portion when sediment is being entrained from the bed; and
- an upper portion when overbank flow occurs and flow width expands rapidly, whereas flow depth increases relatively slowly.

Hydraulic geometry is also used to examine downstream changes in the relations between discharge and channel form. At-a-station hydraulic geometry exponents are typically smaller and more variable than those for downstream hydraulic geometry (DHG) relations, indicating lower rates of change with discharge at a cross section than down a river.

Downstream hydraulic geometry

DHG relations take the same basic form as at-a-station relations, with power functions relating width, depth, and velocity to discharge. DHG relations describe changes in w, d, and v as a discharge of the same frequency varies in magnitude downstream (Leopold and Maddock, 1953). Mean annual or bankfull discharge is typically used for DHG analyses (Park, 1977), with the assumption that channel dimensions primarily reflect the forces exerted by a flow with this recurrence interval. DHG relations are essentially scaling functions for changes in cross-sectional geometry resulting from downstream changes in discharge magnitude.

Typical exponent values for the relations between Q and w, d, and v, respectively, are 0.4–0.5, 0.3–0.4, and 0.1–0.2 (Park, 1977). DHG is used for engineering stable channels under the assumption that $w \sim Q^{0.5}$ when Q has a recurrence interval of 1–2 years. As noted in earlier discussions of bankfull and dominant discharge, this can be a reasonable assumption for channels with limited hydrologic variability, but can be an inappropriate assumption for channels with greater hydrologic variability in which channel width reflects a discharge of longer recurrence interval. DHG is also used to predict the effects of flow regulation, and to understand the geomorphic role of floods or infer the magnitude of past floods based on channel dimensions (Ferguson, 1986) (supplemental Section 3.2.1).

Strong correlations between Q and w, d, and v exist where channel boundaries have relatively low erosional thresholds, and where substantial variations in sediment supply or imposed gradient are not present. Numerous site-specific case studies, however, document weak or poorly developed DHG relations in mountainous regions with persistent geomorphic effects from glaciation or strong downstream contrasts in substrate erodibility or grain-size distribution of sediment inputs (Wohl, 2010).

Wohl (2004b) proposed an empirical threshold for well-developed DHG relations in high-relief drainage basins as a function of the ratio of stream power (hydraulic driving forces) to sediment size (substrate resistance). Higher values of stream power associated with large discharge per unit drainage area, as in the tropics, can result in well-developed DHG relations even in channels with strong colluvial or bedrock influences (Pike et al., 2010). DHG relations for bedrock channels are within the range of those in alluvial channels, although bedrock channels tend to be narrower and deeper (Montgomery and Gran, 2001; Wohl and David, 2008).

Both forms of hydraulic geometry are scaling relations between discharge and the other variables of the continuity equation. The exponents of the hydraulic geometry relations reflect the relative magnitude of response among w, d, and v to changes in Q, and can be used to infer the nature of channel adjustment. The DHG average width exponent of ~0.5, for example, indicates that approximately half of the channel adjustment to downstream increases in discharge occurs via channel widening, whereas the at-a-station average depth exponent of 0.42 indicates that adjustment to greater flows at most cross sections occurs primarily via increased flow depth rather than increased width. Precisely predicting channel response to changing discharge is difficult for at least two reasons:

- interactions among w, d, v and other channel form variables (e.g., bedform amplitude) not included in hydraulic geometry relations; and
- the influence of sediment supply, which is also not included in hydraulic geometry relations.

At-a-station and downstream hydraulic geometry are nonetheless useful in understanding the manner in which channel form adjusts to fluctuating discharge.

5.1.4 Lane's balance

Lane's balance refers to a conceptual model of channel adjustment that includes several cross-sectional parameters and accounts for both water and sediment discharge. Lane (1955) conceptualized equilibrium within a channel segment as reflecting a balance among discharge (Q_w), channel gradient (S), sediment load (Q_s), and sediment size (D_s)

$$Q_w S \propto Q_s D_s \tag{5.4}$$

This relation is referred to as *Lane's balance* and has been widely depicted using a drawing of a balance (Figure 5.3) (Grant et al., 2013). This drawing is intuitively appealing and easy to understand. An increase in sediment load ($\uparrow Q_s$), for example, will cause aggradation, coarsening ($\uparrow D_s$), and steepening ($\uparrow S$) of the stream.

The ability of Lane's balance to describe river adjustments is inherently limited, however, because the expression does not account for changes in cross-sectional, planform, and bedform geometry that are commonly associated with channel adjustments to changes in discharge and sediment load. Dust and Wohl (2012b) expanded Equation 5.4 to include these terms

$$Q_w \left(\frac{\Delta z}{P\bar{H}_a}\right) \propto Q_s D_s \left(\frac{w}{d}\right) \tag{5.5}$$

where S is proportional to total change in elevation along a channel (Δz) and inversely proportional

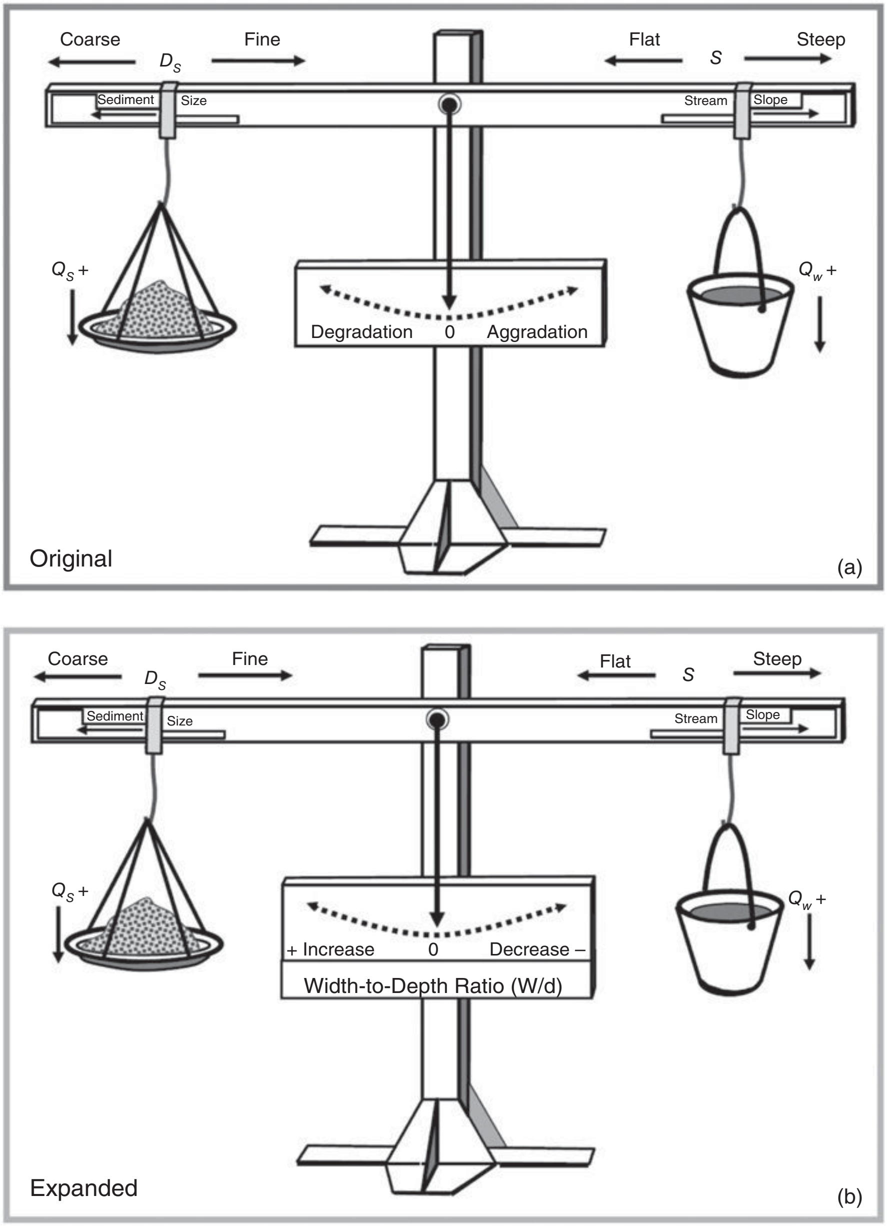


Figure 5.3 An illustration of Lane's relation, which is widely used to conceptualize channel adjustment in response to changes in water or sediment yield. (a) The original illustration, based on a drawing by Whitney Borland of Colorado State University. (From Dust and Wohl, 2012b, Figure 1.) (b) An illustration of an expanded version of Lane's relation, including adjustments to channel width-to-depth ratio. (From Dust and Wohl, 2012b, Figure 9.)

to sinuosity (P) and bedform amplitude ($\bar{H}_a$), and (w/d) is width/depth ratio. Equation 5.5 suggests that an increase in sediment load ($\uparrow Q_s$) will cause aggradation ($\uparrow z$), decreased sinuosity ($\downarrow P$) and bedform amplitude ($\downarrow \bar{H}_a$), and decreased width to depth ratio ($\downarrow w/d$), as well as coarsening and steepening.

Either version of Lane's balance is useful primarily as a conceptualization of the various potential ways in which channel geometry can adjust to the

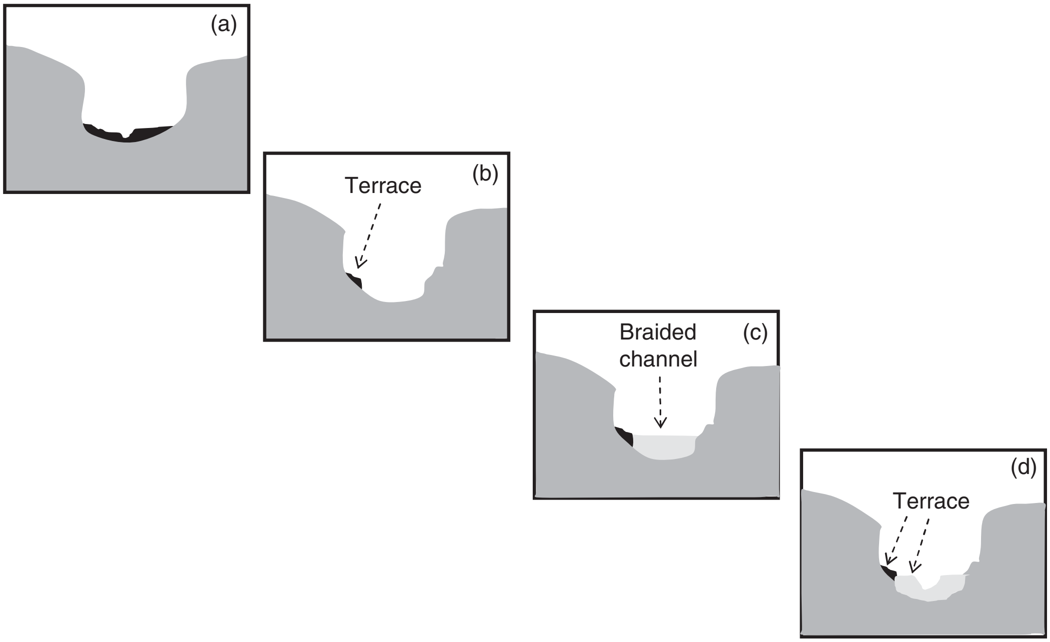


Figure 5.4 Complex response of a channel to a single fall in base level: perspective is upstream within the valley (gray shading) at a point midway along the channel between the mouth and the headwaters. (a) Stream alluvium (black) deposited before base level lowering. (b) After base level lowering, a knickpoint migrates upstream, causing the channel to incise into alluvium and bedrock of the valley and leaving the alluvium as a terrace. Following incision, bank erosion widens channel and partially destroys the terrace. (c) Continued upstream migration of the knickpoint causes increased sediment discharge to mid-portion of river. Inset alluvial fill is deposited as sediment discharge from upstream increases. A broad, shallow, braided channel (lighter gray shading) develops. (d) Upstream migration of the knickpoint ceases and the sediment supply to downstream channel segments declines. The channel becomes deep and narrow and incises slightly, creating a second terrace. (From Schumm and Parker, 1973, Figure 1.)

controlling variables of Q_w and Q_s. An analogous method to quantitatively predict channel adjustment remains elusive because of the numerous interdependent variables.

5.1.5 Complex response

Schumm (1973) and Schumm and Parker (1973) introduced the phrase *complex response* to describe asynchronous, discontinuous river response to a single external perturbation. During flume experiments, Schumm and Parker (1973) observed that base level fall initiated a head cut that migrated upstream. Upstream migration increased sediment load to channel segments downstream from the head cut, which then began to aggrade (Figure 5.4). Sediment supply declined once the head cut reached its furthest upstream point, causing downstream reaches to incise, and sometimes creating a second head cut that migrated upstream. Consequently, the upstream portion of a river can be incising while the downstream portion is aggrading. Any given point within the channel alternates between incision and aggradation with time, sometimes going through multiple cycles of incision/aggradation in response to the initial base level fall before the channel once more stabilizes.

Subsequent studies have applied the idea of complex response to a variety of spatial and temporal scales and types of rivers (Marston et al., 2005; Harvey, 2007). In general, although a river can incise or aggrade throughout its length, the more common scenario is that aggradation in some part of the river's length is likely associated with enhanced erosion of the channel boundaries elsewhere. Among the implications of complex response is that

- multiple terraces along a river may not represent multiple external changes (Womack and Schumm, 1977) (Section 6.2);
- incision or aggradation within a segment of channel can be a transient response to upstream or downstream changes in channel geometry, base level and sediment supply, with the time span of the transience dependent on factors such as the size of the channel and the erodibility of the bed and banks; and
- consequently, restoration or management that seeks to limit incision or aggradation may actually enhance undesirable channel change if implemented just as the direction of channel response (i.e., incision vs. aggradation) is changing.

5.1.6 Channel evolution models

The ideas underlying Lane's balance and complex response, in particular, are incorporated into *channel evolution models*. Alluvial channels can develop very small *w/d* ratios when undergoing rapid incision, but this is typically a transient condition. Alluvial channels in diverse environments that incise in response to base level fall or changes in water and sediment supplied to the channel pass through a characteristic sequence of channel geometry with time that is summarized in channel evolution models. The channel adjustment described in these models typically begins with a relatively narrow, deep channel. The channel subsequently widens, aggrades, and eventually stabilizes. Schumm et al. (1984) proposed empirical relations (e.g., top width = 46.77 (drainage area)$^{0.39}$) between top width and drainage area, derived from observations on stabilized channels, which can be used to identify whether incising channels are close to a stability threshold at which widening stops. A specific form of this relation can be useful within a limited geographic region, but the coefficient and exponent vary between regions.

Schumm et al. (1984) initially proposed a five-stage model of channel evolution. This has been modified to six or more stages (Simon and Rinaldi, 2013) that reflect shifts in the dominant adjustments and associated rates of sediment transport, bank stability, and cross-sectional geometry (Figure 5.5). The time necessary to develop each stage of evolution and the relative magnitudes of vertical and lateral adjustments vary widely between different streams and stream segments (Simon and Rinaldi, 2006), and the entire sequence can require 10^2–10^3 years (Simon and Castro, 2003) to complete. Like Lane's balance and complex response, channel evolution models are useful conceptualizations rather than quantitative predictions.

Incised channels of the type described in channel evolution models can form anywhere, but are particularly common in arid and semiarid regions. Channels in these regions undergo repeated episodes of alternating incision and aggradation in response to internal thresholds, as described in Section 1.5.1 (Schumm and Hadley, 1957) or to external changes in flood magnitude and frequency (Webb and Baker, 1987), other aspects of climate (Leopold, 1976), or land uses such as grazing and groundwater withdrawal (Cooke and Reeves, 1976). Artificially channelized streams, which are described in Chapter 7, also typically go through the stages described in channel evolution models following completion of channelization.

Adjustments to cross-sectional geometry represent an intermediate level of channel response to changing water and sediment dynamics. Bed grain-size distribution and bedform configuration are more likely to change first as water and sediment supplies fluctuate, as described in the preceding chapter. Adjustments to channel *w/d* ratio typically represent the next level, in response to larger magnitude or more prolonged fluctuations in water and sediment. Changes in channel planform represent a third level of channel adjustment.

5.2 Channel planform

Several classification schemes have been proposed to describe the wide variety of forms assumed by rivers when viewed on a two-dimensional planar

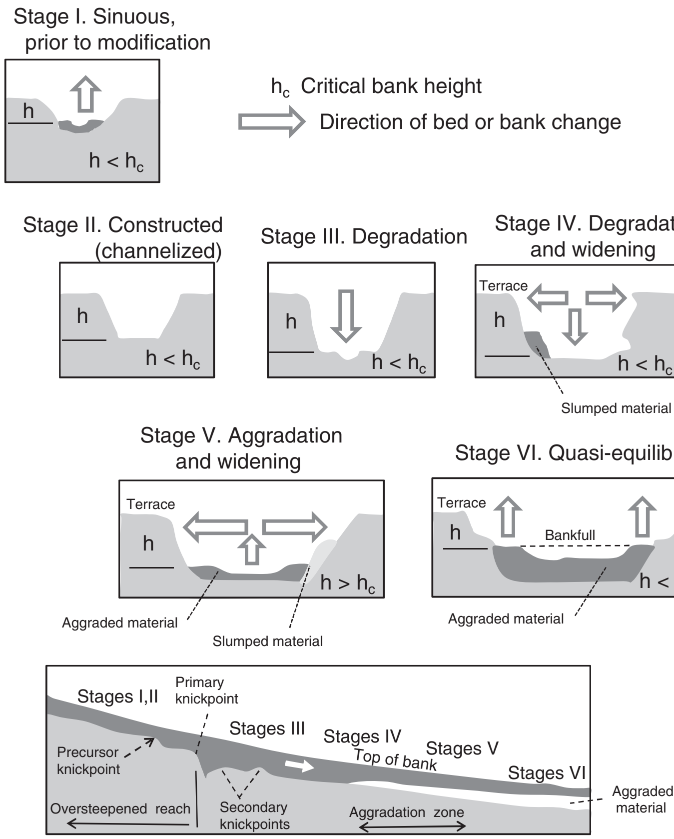


Figure 5.5 Illustration of a six-stage channel evolution model. View is upstream or downstream in the first six boxes. The lower box contains a longitudinal profile illustrating different stages occurring simultaneously along different segments of a channel. Light gray shading is valley material (sediment or bedrock), darker gray is recent alluvium. (From Simon and Castro, 2003, Figure 11.11, and Wohl, 2010, Figure 4.14.)

surface such as a map (Buffington and Montgomery, 2013). An obvious distinction is between rivers with single channels and those with multiple channels. Single channels are typically differentiated on the basis of *sinuosity*, the ratio defined by actual flow path downstream to straight line distance between two points. Multiple channels are typically differentiated based on the lateral mobility of secondary channels.

Leopold and Wolman (1957) proposed a tripartite classification of straight, meandering, and braided channels. Although this classification is still used, there are many channel planforms that do not fit well within these categories. Schumm (1985) proposed a broader range of 14 channel types (Figure 5.6), and other investigators have described additional categories such as wandering gravel-bed rivers (Carson, 1984), anabranching (Brice et al., 1978), and compound rivers that regularly alternate between two or more planforms over time.

Any classification imposes more or less arbitrary boundaries on a continuum of channel form. Ideally, the boundaries reflect thresholds in the processes that create channel form. As with substrate, bedforms, and cross-sectional geometry, channel planform reflects an adjustment of the rate of energy expenditure in relation to water and sediment supplied to the channel and erosional resistance of the channel boundaries. Any channel planform that deviates from a single, straight channel—as many planforms do—represents an adjustment of resistance to flow. A channel with meanders, for example,

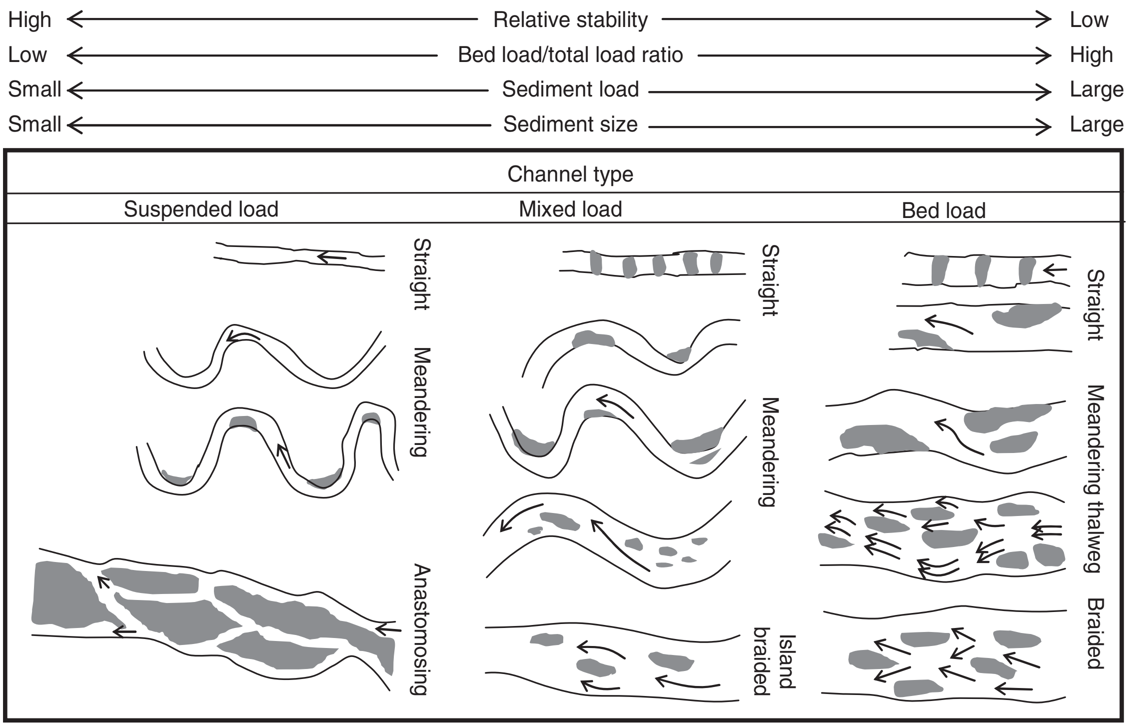


Figure 5.6 Classification of alluvial channel pattern (from Schumm, 1981). Gray shading indicates depositional areas in the form of islands, bars or riffles. Arrows indicate flow paths.

effectively has a longer flow path for a given downstream length and gradient, and thus greater boundary resistance than an otherwise equivalent straight channel. The most commonly occurring planforms are described in more detail next.

5.2.1 Straight channels

Straight channels have a single channel with sinuosity less than 1.5. Straight channels can be straight because they are closely confined by steep valley walls or other uplands such as terraces, or they can occur in erodible, alluvial boundaries. Even straight channels in erodible alluvial boundaries, however, remain straight because the banks have sufficient erosional resistance relative to available flow energy to limit bank erosion (Paola, 2001). Straight channels can be further distinguished as those with or without mobile bedforms, those with alternating bars and a sinuous thalweg, and slightly sinuous channels with point bars.

Straight alluvial channels without mobile bedforms have mostly suspended sediment, a low gradient, and a deep, narrow cross section. Such channels are rare (Schumm, 1985). Straight alluvial channels with mobile bedforms are typically sand-bed channels in which the sequence of immobile bed, ripples, dunes, mobile plane bed, and antidunes described in Section 4.3.1 occurs as flow changes.

The majority of straight alluvial channels in mixed substrates or substrates coarser than sand have pool–riffle sequences of greater or lesser mobility, depending on the substrate and sediment supply. The downstream alternation between pools and riffles is associated with cross-sectional asymmetry and undulations in bed elevation, so that a regular channel planform does not equate to regularity in other dimensions, or to uniformity of boundary roughness. Pools typically occur on alternate sides of the channel in a downstream direction, with intervening depositional features such as riffles and alternate bars (Figure 5.7). The *thalweg*—the line of

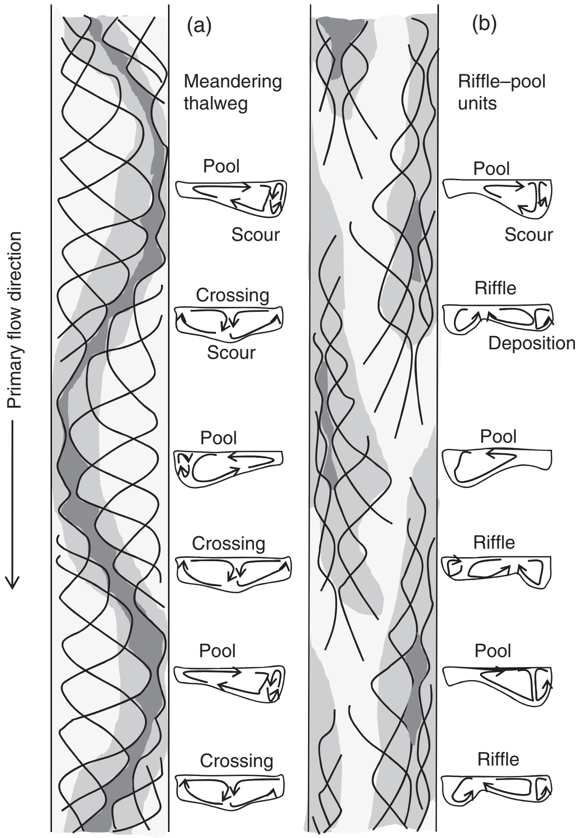


Figure 5.7 Plan view showing a sinuous thalweg and alternate bars within a straight channel, with associated helical flow and cross-sectional asymmetry shown by small channel cross-sectional views at right. (a) Model of twin, periodically reversing, surface-convergent helical cells based on work by Einstein and Shen (1964). (b) Model of surface-convergent flow produced by interactions between the flow and a mobile bed, creating pool–riffle units of alternate asymmetry based on work by Thompson (1986). Black lines indicate currents and progressively darker gray shading indicates progressively deeper portions of the channel (from Knighton, 1998, fig 5.15, p. 199.)

deepest flow—is thus sinuous even though the channel boundaries are straight. The flow alternately converges passing through pools and diverges passing over depositional zones.

The ubiquity of either a sinuous thalweg or a sinuous channel planform presumably reflects an inherent tendency in water flowing over a rough surface to develop a rotational component of flow as a result of greater hydraulic resistance along the channel boundaries than in the center of the channel. This rotational component, described as *helical flow*, creates downstream alternations in the location of greatest velocity that are expressed in differences in boundary erosion, sediment transport and deposition, and channel geometry. Sinuosity in flow and channel planform scales consistently across features as diverse as channels less than a meter wide on glacial ice and channels more than a kilometer wide in very large rivers (Leopold, 1994).

5.2.2 Meandering channels

Single channels with a sinuosity greater than 1.5 appear to be the most widespread and common type of channel planform (Leopold, 1994). Meandering channels can be differentiated based on the regularity of bends (Figure S5.2) (Kellerhals et al., 1976) into

- *irregular meanders* with a weakly repeated downstream pattern;
- *regular meanders* with a repeated pattern and a maximum deviation angle between the channel and down-valley axis <90 degrees; and
- *tortuous meanders* with a less clearly repeated pattern and a maximum deviation angle >90 degrees.

This differentiation reflects the fact that naturally formed meanders are seldom perfectly regular, but instead include randomness that reflects local controls on the erodibility of the channel boundaries.

In any given meander, the outer banks are commonly steep and eroding and a pool is present at the bend apex. Cross-sectional bed topography slopes downward from the point bar on the opposite inner bank. Riffles are present in the inflection regions of the bend, in the straight limbs between bend apices, where cross-sectional and bank geometry are more symmetrical (Hooke, 2013).

Meander geometry is typically characterized using meander wavelength, λ, and radius of curvature, r_c, for individual bends. Path direction, θ,

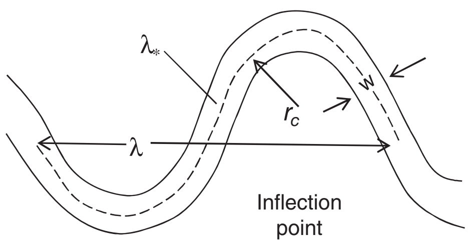


Figure 5.8 Components of meander geometry.

and change of direction, $\Delta\theta$, characterize meander geometry for multiple bends (Figure 5.8).

Langbein and Leopold (1966) introduced the idea of modeling meander geometry using the equation for a sine-generated curve

$$\theta = \omega \sin kx \tag{5.6}$$

where θ is channel direction expressed as a sinusoidal function of distance x, with parameters ω for the maximum angle between a channel segment and the mean down-valley axis and $k = 2\pi/\lambda$. A sine-generated curve represents a path in which the sum of squares of changes in channel direction per unit length is minimized, which effectively distributes stress uniformly along the curve. A sine-generated curve is a good approximation of the geometry of regular, symmetrical meanders (Leopold, 1994). Many—perhaps the majority of—natural meanders, however, are not symmetrical (Carson and Lapointe, 1983). Measurements of meander geometry across a range of environments suggest that λ is typically 10–14 times the channel width and r_c is typically 2–3 times the channel width (Knighton, 1998).

Just as channel width is approximately proportional to $Q^{0.5}$, meander wavelength also typically varies as $\sim Q^{0.5}$, whether Q is mean annual flood, mean annual discharge, mean monthly maximum discharge, or some other measure of flow that is likely to transport most of the sediment and thus strongly influence channel form parameters (Dury, 1965; Knighton, 1998). Schumm (1963, 1967) demonstrated that, for a given discharge, meander wavelength also varies with boundary cohesion and gradient. Wavelength decreases as boundary cohesion and gradient increase.

The thalweg of a meandering river does not maintain a central location along the channel in a downstream progression. Instead, the thalweg migrates across the channel through each bend from near the inner bank at the bend entrance to near the outer bank at the bend exit (Figure 5.9). The strongly helical flow in a meandering channel is expressed as superelevation of the water surface against the outer bank of each bend and a transverse current directed toward the outer bank at the surface and toward the inner bank at the bed. Helical flow facilitates preferential erosion of the outer bank and deposition on a point bar along the inner bank.

The details of strength and location of the transverse, secondary currents reflect flow stage, meander geometry, and channel cross-sectional geometry. Secondary currents weaken with increasing stage and the flow follows a straighter path. Tighter bends with a lower ratio of radius of curvature to channel width (r_c/w) have stronger secondary circulation. A large width/depth ratio is associated with more extensive development of point bars, and point bars also reflect grain-size distribution (Hooke, 2013).

Although models, in particular, typically assume relatively symmetrical and regular meander bend geometry, natural meanders can have substantial deviations in form and flow resistance, with associated deviations from an idealized distribution of hydraulic variables. The presence of instream wood along a meander, for example, strongly modifies the three-dimensional (3D) flow structure in a manner highly dependent on the arrangement, density, and mobility of the wood (Daniels and Rhoads, 2004). In some bends, wood can create sufficient resistance to reduce flow velocity and constrain the high-velocity flow and helical motion to the channel center, whereas in other bends the wood can strongly deflect the flow away from the banks (Daniels and Rhoads, 2004).

Hypotheses regarding the initiation of meandering tend to focus either on inherent properties of flow, or on interactions between the flow and a

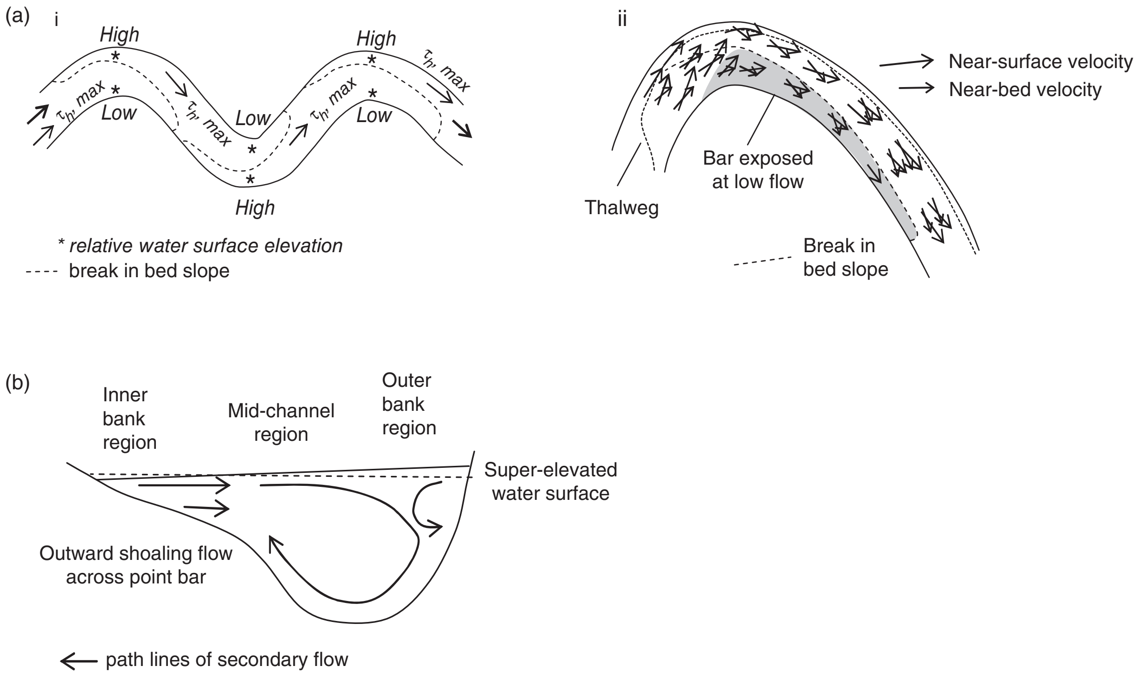


Figure 5.9 Sinuosity of the thalweg and associated velocity and shear stress in meandering channels. (a) (i) Location of maximum boundary shear stress (τ_b) and (ii) flow field in a bend with a well-developed bar (from Dietrich, 1987). (b) Secondary flow at a bend apex showing the outer bank cell and shoaling-induced outward flow over the point bar (from Markham and Thorne, 1992).

mobile channel boundary. The argument for flow properties rests on the fact that water flowing down a rough, inclined surface—even a relatively smooth plate in a flume—in the absence of sediment, develops sinuosity at the base of the flow where frictional resistance is greatest. Progressively more of the flow depth is involved in sinuous flow as discharge and velocity increase. The forces involved in this type of flow do not necessarily scale well to natural channels (Hooke, 2013), however, which typically have at least partly mobile beds. Arguments for interactions between the flow and the boundary as the underlying cause of meandering emphasize deformation of the bed leading to the development of alternate bars that initiate meandering by deflecting flow toward the opposite banks (Seminara and Tubino, 1989). Sinuous channels develop in materials without bars, however, such as those formed atop glacial ice or in bedrock.

Most likely, meanders are not the outcome of a single cause. As explained by Güneralp and Marston (2012), helicoidal, *curvature-driven secondary flow* is generated by superelevation of the water surface at the outside of a bend. This creates substantial cross-stream variation in velocity, which redistributes downstream momentum, leading to a decrease in bed shear stress and deposition of a point bar along the inner bend, as well as a downstream increase in bed shear stress and erosion along the outer bend. The point bar deflects flow laterally toward the cut bank, creating *topographically driven secondary flow*. Seminara (2006), Güneralp and Marston (2012), and Hooke (2013) reviewed theoretical and mechanical explanations for meandering in more detail.

Attention has also focused on why meanders form. Extremal hypotheses focus on the influence of meanders on energy expenditure. The reasoning is that meanders either (i) create the most uniform rate of energy expenditure along a channel by minimizing variance in hydraulic variables (Langbein and Leopold, 1966), or (ii) establish a minimum channel slope for given input conditions (Chang, 1988). Subsequent research, however, emphasizes the absence of equilibrium (Hooke, 2013), focusing instead on continuous evolution and instability (Eaton et al., 2006). Discharge, sediment size and supply, bank resistance, and gradient are still acknowledged to influence meander migration and morphology, implying that any change in these variables results in a response in meander form and process (Hooke, 2013).

Meanders can migrate very slowly, but even deeply incised bedrock channels that are sinuous show evidence of meander migration during incision (Harden, 1990). Migration of individual meanders (Daniel, 1971) can occur through

- translation, which is downstream shifting of the bend without alteration in basic shape;
- extension, during which the bend increases its amplitude by migrating across the valley;
- rotation, in which the bend axis changes orientation; and
- lobing and compound growth, during which the bend becomes less regular and symmetrical.

Individual bends along a river can have different styles and rates of migration, and bends typically deform and become asymmetrical with migration. The fastest migration tends to occur when r_c/w is between 2 and 3 (Hickin and Nanson, 1984). Ratios outside of this range lead to preferential migration in the upstream or downstream limb of the bend, with associated changes in curvature rather than rapid migration of the entire bend (Knighton, 1998).

Meander migration that increases the amplitude and tightness of bends can exceed a stability threshold and trigger a *chute cutoff* that creates a shorter channel across the inside of the point bar or a *neck cutoff* at the base of the bend (Constantine et al., 2010). Because a cutoff increases channel gradient and local transport energy, the cutoff becomes the main channel and the longer flow path becomes a secondary or overflow channel that accumulates sediment during high flows. Secondary channels can persist for decades to centuries, depending on the rates of sediment filling (Figure S5.3). While present, secondary channels can form floodplain wetlands and increase habitat diversity for aquatic and riparian organisms. Secondary channels gradually fill with sediment settling from suspension during overbank flows. Because this sediment is commonly finer grained than adjacent depositional areas of the floodplain, even secondary channels that completely filled with sediment decades or centuries earlier create persistent differences in groundwater, soil moisture, and plant communities.

Supplemental Section 5.2.2 discusses measurement and modeling of meandering rivers.

5.2.3 Wandering channels

Wandering gravel-bed rivers have rapid bend migration, numerous bars, and frequent dissection of point bars. Wandering rivers are probably transitional between meandering and braided planforms (Carson, 1984) (Figure S5.4). Wandering channels have also been subdivided into those with single channels, high channel migration rates, and frequent dissection of point bars, and those with multiple channels, a large supply of bed sediment, and low to moderate bank erodibility (Carson, 1984). Wandering channels are most commonly described for mountainous regions with headwater glaciers or downstream from large terraces, although this channel planform can occur in any environment (Church, 1983; Burge, 2005). Although the number of papers using the category of wandering channels is limited, the designation of this channel type does reflect the fact that distinguishing meandering, braided, and anabranching channels is not always straightforward (Nanson and Knighton, 1996).

5.2.4 Braided channels

Braided rivers are multi-thread channels in which flow is separated by bars within a defined channel (Figure S5.5). Some of the bars can be submerged at high flows, but all are typically exposed at low flows. The degree to which bars are stabilized by vegetation varies, but the usual distinction between multi-thread braided rivers and multi-thread anabranching rivers is that bars in braided rivers have less vegetation, are narrow relative to the width of the channel, and are relatively mobile compared to anabranching rivers. When bars and islands are distinguished, mid-channel bars are unvegetated and submerged at bankfull stage, whereas islands are vegetated and emergent at bankfull stage (Brice, 1964). Bars can be:

- longitudinal and formed of crudely bedded gravel sheets;
- linguoid bars that are lobate in shape, of sand or gravel deposited by downstream avalanche-face progradation; and
- point or side bars formed by coalescence of smaller bedforms such as dunes and linguoid bars at sites of lower energy (Miall, 1977) (Figure S5.6).

Flume experiments with initially straight channels indicate that braiding can develop from the formation of single alternate bars, single mid-channel bars, or multiple mid-channel bars (Ashmore, 2013). These bars are low-amplitude bedforms occupying most of the channel width and connected to upstream scour pools. The bars deflect flow and start the development of channel sinuosity. Braiding develops either by cutoff of single bars at a critical bend amplitude, or bifurcation (splitting of flow) around mid-channel bars (Ashmore, 2013). Braiding is maintained by repetition of these initial bar-scale processes within the individual channels of the braided network, but the initially simple, well-defined pool-bar units (pool head to downstream bar margin) of a single channel are replaced by the confluence-bar/bifurcation units (the distance over which two channels join and then split again downstream) that form a basic morphological element of braided rivers (Bridge, 2003; Ashmore, 2013) (Figure 5.10).

Repeated division and joining of channels, and associated divergence and convergence of flow, correspond to rapid shifts in channel position and the size and number of bars, particularly during floods. Braiding is produced by processes active at higher flows, rather than resulting solely from dissection of bars during low flows (Ashmore, 2013). Flume experiments indicate a strong, positive relationship between sediment supply and frequency of channel avulsion (lateral movement and formation of a new channel), such that increased sediment supply causes greater channel mobility, bifurcation, and avulsion (Ashworth et al., 2007). Bars more strongly influence bifurcation processes as the bar height above the bed (bar amplitude) increases (Bertoldi et al., 2009). Adjustments to braided channel geometry alter flow resistance, and the degree of braiding tends to increase with slope (Parker, 1976).

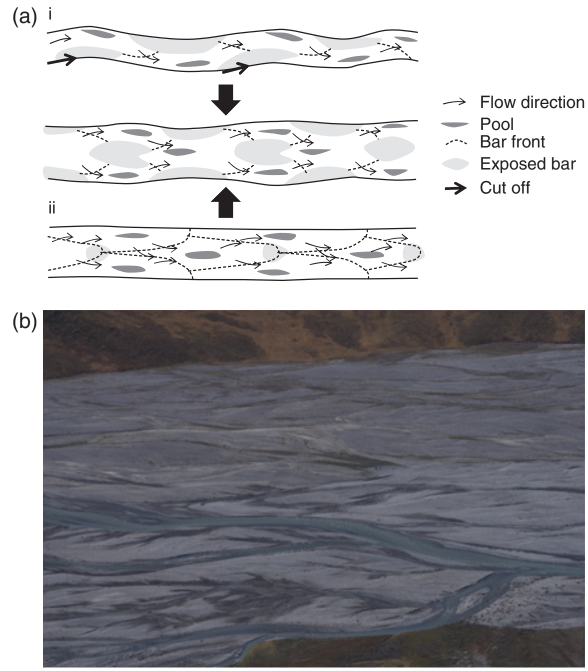


Figure 5.10 (a) Schematic illustration of the development of braiding from initial (i) alternate bars by channel widening and chute cutoff, or (ii) central or higher mode bars. (From Ashmore, 2013, Fig 3.) (b) Photograph of a wide (~1 km) braided river in northern Alaska. Flow is from left to right.

The degree of braiding has been quantified using various braiding indices. The most common approach is to count the mean number of active channels or braid bars per transect across the channel belt (Bridge, 2003; Egozi and Ashmore, 2008).

Braided channels are much less common than meandering channels. However, both the rock record prior to the evolution of land plants

(Montgomery et al., 2003a; Davies and Gibling, 2011) and flume experiments suggest that braided rivers are the default channel planform in rivers lacking sufficient riparian vegetation or cohesive bank sediments to substantially increase bank resistance to erosion (Paola, 2001). Where woody riparian vegetation expands, commonly as a result of changes in flow regime, braided channels can metamorphose to a meandering or anabranching planform (Nadler and Schumm, 1981; Piégay and Salvador, 1997). Field studies indicate that the rate and locations of growth of woody riparian vegetation strongly influence the formation and erosional resistance of islands (Gurnell and Petts, 2006), as well as the stability of outer banks along a braided channel. Instream wood can also strongly influence the location and stability of bars, as well as the establishment of riparian forests on the bars, when sediment is deposited around a bar-apex logjam (Abbe and Montgomery, 2003) (Figure S5.7). The tendency toward braiding may be influenced by a river's ability to turn over its bed within the characteristic time for riparian vegetation to establish and grow to a mature, scour-resistant state (Paola, 2001). This dimensionless time-scale parameter can predict whether a channel will braid (Hicks et al., 2008).

Braided channels occur in diverse environments and across a broad range of scales, and are particularly common in arid and semiarid regions, downstream from glaciers, and in mountainous environments with abundant coarse sediment supply and limited riparian vegetation (Ashmore, 2013) (Figure S5.8). Proglacial braided channels are known by the Icelandic word *sandurs* at the point where the channel system expands freely, and *valley sandurs* where development of the channel network is confined by valley walls (Krigstrom, 1962).

Braiding tends to be associated with four conditions, although no single one of these conditions is either sufficient or necessary to create a braided channel (Knighton, 1998).

1. Abundant bedload can cause braiding if the channel lacks capacity to transport the volume of sediment supplied, or competence to move the size of sediment supplied (Griffiths, 1979). Locally reduced transport capacity can facilitate sediment deposition that allows a bar to form and grow, deflecting the current toward the adjacent banks, creating local bank erosion, and introducing more sediment into the flow. This is the *central bar mechanism* that Leopold and Wolman (1957) invoked to explain the initiation of braiding. Related to this is the *transverse bar conversion mechanism* of initiating braiding, in which flow convergence through a pool scours the bed and provides sufficient bedload for deposition downstream from the pool where the flow diverges, eventually causing flow to be deflected around an elevated bar (Ashmore, 1991). Increased bedload supply causes aggradation and an increase in the degree of braiding, whereas decreased supply has the opposite effect (Germanoski and Schumm, 1993; Thomas et al., 2007).
2. Erodible banks facilitate continued channel widening and the development of multiple bars in wide, shallow flow with heterogeneous transport capacity. This corresponds to the argument presented earlier: single-thread or anabranching channels tend to form where cohesive banks result from silt and clay or from riparian vegetation.
3. Rapid fluctuations in discharge contribute to bank erosion and heterogeneous bedload movement, and large floods can initiate braiding, in part by removing stabilizing riparian vegetation and dramatically increasing bank erosion and channel width (Burkham, 1972; Friedman and Lee, 2002; Jaquette et al., 2005) (Figure S5.9). Some types of channels alternate repeatedly between braided (immediately after a large flood) and meandering (developing gradually during lower discharges over years to decades following a large flood).
4. Steep valley gradients appear to promote braiding, although high stream power (QS) may be a better measure than simply gradient (Knighton, 1998). Different threshold values have been proposed, but braided rivers tend to have higher values of stream power than meandering rivers of similar grain size (Ashmore, 2013). The physical

explanation for this observed correlation remains under debate.

In addition to mechanisms of initiating braiding through primarily depositional processes, as described earlier, braiding may also initiate in response to erosion of bars. Flume experiments indicate that dissection or the formation of cutoff channels across various types of bars occurs when flow follows a steeper route across a bar, causing headward incision across the bar, which increases the braiding index (Ashmore, 1991; Germanoski and Schumm, 1993). The presence of bars is still basic to developing a braided channel, indicating that the fundamental underlying process is local deposition in response to loss of transport capacity.

Once braiding starts, all of the erosional and depositional processes described earlier may be operating to maintain braiding, as long as short-term transience in bedload transport is maintained. Such transience is closely tied to longitudinally alternating convergent flow zones at confluences, in which sediment from upstream is transported and bed scour occurs, and divergent flow zones at bifurcations, in which sediment is deposited (Ashmore, 2013).

Confluence geometry reflects relative discharge, junction angle, and orientation of the confluent channels. Maximum depth of scour increases as the junction angle increases and as discharges in the confluent channels approach equal magnitude (Mosley, 1976; Best and Rhoads, 2008). *Avulsion frequency* is the frequency at which a new channel forms that carries ≥50% of the discharge of the old channel. Avulsion frequency scales with the time necessary for bed sedimentation to produce a deposit equal to one channel depth, at which time the channel is likely to avulse. This is reflected in a dimensionless mobility number based on the relative rates of bank erosion and channel sedimentation (Jerolmack and Mohrig, 2007).

As in the case of sinuosity in meandering channels, braiding represents a morphological adjustment of channel gradient to a stable configuration: in this case, by progressive channel subdivision, which effectively alters the channel gradient by altering length of flow path for a given vertical drop. Mid-channel bars exhibit similar width/length ratios over a wide range of spatial scales (Kelly, 2006). This suggests that braiding is a self-organized, emergent property of the interaction between flow and a non-cohesive sediment bed that must be constrained in some way if morphology other than braiding is to develop (Paola, 2001; Ashmore, 2013).

Supplemental Section 5.2.4 discusses measurement and modeling of braided rivers.

5.2.5 Anabranching channels

Anabranching channels are multi-thread channels in which individual channels are separated by vegetated or otherwise stable bars and islands that are broad and long relative to the width of the channels, and that divide flows at discharges up to bankfull (Nanson, 2013). The islands persist for decades to centuries and are similar in elevation to the floodplain (Knighton, 1998). Individual anabranching channels can be straight, meandering, or braided but, unlike distributary networks, the channels in an anabranching network eventually rejoin. Anabranching channels are relatively uncommon, but occur in very diverse environments (Figure S5.10).

Anabranching occurs in bedrock channels and the characteristics of individual channels appear to be influenced by joint geometry in the bedrock. Anabranching reaches are particularly common immediately upstream of waterfalls (Kale et al., 1996; Tooth and McCarthy, 2004), but can also occur in lower gradient reaches without knickpoints (Heritage et al., 2001).

Anabranching is also particularly common in very large alluvial rivers (Jansen and Nanson, 2004; Latrubesse, 2008), although many of the historically anabranching segments of these rivers have been channelized to single-thread channels (Pišút, 2002). Anabranching is the dominant channel pattern of the Amazon, Congo, Orinoco, Parana, and Brahmaputra, among other large rivers, and anabranching was historically present along rivers such as the Danube and Rhine (Latrubesse, 2008).

Alluvial anabranches can develop as erosional channels scour into the floodplain during channel avulsion. Avulsion can also be triggered by sediment accumulation within a channel, particularly where an obstruction such as a channel-spanning logjam creates a backwater effect (Abbe and Montgomery, 2003; O'Connor et al., 2003; Wohl, 2011c). Rapid aggradation can produce frequent avulsions and a network of channels in various stages of formation and abandonment that can be wandering or anabranching (Nanson, 2013). Anabranches can also develop from mid-channel bars that become islands dividing the flow in previously wider channels, particularly in very low gradient channels. Finally, anabranches can develop from delta progradation and modification of the distributary network (Nanson, 2013). In this situation, anabranching may reflect an efficient means of redistributing and storing excess sediment across a wide valley, because the lower *w/d* ratios within individual anabranches enhance bed shear stress and sediment transport (Huang and Nanson, 2000). Conversely, anabranching increases boundary resistance and may effectively consume surplus energy along rivers with very low sediment loads (Nanson and Huang, 2008).

Although anabranching and anastomosing are sometimes used synonymously, Nanson and Knighton (1996) distinguished six types of anabranching rivers, including anastomosing.

1. *Anastomosing rivers* form in cohesive sediment and low gradients, and have very low stream power. This type of river has been described in diverse environments including the Rocky Mountains of Canada, arid Central Australia, Africa, and Wyoming, USA, and tropical South America.
2. Sand-dominated, island-forming anabranching rivers rely on riparian vegetation to provide bank cohesion. Such channels have been described for Australia, and have relatively low values of stream power.
3. Mixed load, laterally active anabranching rivers have moderate values of stream power. This type of anabranching river ranges from small channels in Australia to very large rivers in South America and Asia.
4. Sand-dominated, ridge-forming anabranching rivers have long, narrow, parallel ridges stabilized by riparian vegetation and straight anabranches with steep banks. This form of anabranching also has moderate stream power and has mostly been described in Australia.
5. Gravel-dominated, laterally active anabranching rivers are wandering channels with higher stream power. Although the basal coarse sediment is overlain by finer sediment on the islands, vegetation provides most of the bank stability. These channels, in particular, can be facilitated by obstructions such as logjams.
6. Gravel-dominated, stable anabranching rivers also exhibit higher stream power. Like type 5 channels, riparian vegetation is critical to enhancing bank stability in these channels and obstructions appear to facilitate anabranching. The steeper, confined valleys in which type 6 channels occur limit lateral mobility.

5.2.6 Compound channels

Compound channel can be used to describe any channel that has distinct low- and high-flow portions, including channels that simply overflow onto a floodplain. Compound channel can also refer specifically to low- and high-flow portions with distinctly different planforms. Examples of the latter include proglacial streams (Fahnestock, 1963) or tropical rivers (Gupta and Dutt, 1989) that switch seasonally between meandering and braided as discharge fluctuates.

An intriguing example of a compound stream is Cooper Creek in central Australia. This creek has a clay-bed anastomosing planform at low flow, then dries out completely, allowing the clay particles to aggregate into sand-sized pellets. At the start of the wet season, the pellets are transported as bedload in braided rivers, but the pellets disaggregate as flow continues, transitioning to the anastomosing clay rivers as flows recede (Nanson et al., 1986).

5.2.7 Karst channels

Channels developed in karst terrains form a unique subset of rivers in that they exhibit the properties of open-channel flow, but exist in subterranean environments. As noted earlier, karst processes and forms are associated with rocks that are readily soluble under surface or near-surface conditions, typically carbonates and evaporates. Rivers in karst terrains can flow underground for substantial distances and then abruptly resurface as a spring where the water table and the karst aquifer intersect the surface, or as a river in a pocket valley where an impermeable substrate forces flow to the surface. Conversely, surface streams can abruptly disappear underground in a *blind valley*. Or, a surface stream in a *dry valley* can contain flow only during large runoff inputs, despite existing in a wet climate, because only large runoff inputs effectively fill subsurface conduits and force flow to the surface (Ritter et al., 2011).

Subterranean rivers in cave systems can behave similarly to surface rivers in the sense of transporting sediment from clay to boulder size clasts and developing channel geometry that reflects adjustments between available energy and substrate resistance (Springer et al., 2003). Karst rivers differ from most surface streams formed in bedrock in that a large portion of the sediment load can be carried in solution, and erosion of resistant channel boundaries can produce distinctive sculpted forms such as scallops and pockets via both abrasion and solution (Springer and Wohl, 2002). Along surface channels, a floodplain can attenuate the greater discharge and energy present during a flood. Subterranean karst rivers may respond to enhanced discharge with much greater flow depths and velocities, if the cave passage occupied by river channel is sufficiently large, and may also develop conditions of pipe flow (Springer, 2004). (Pipe flow does not have the free surface found in open-channel flow.) In many respects, subterranean karst channels are analogous to laterally confined bedrock channels at the surface, with downstream changes in substrate erodibility strongly influencing channel geometry.

5.2.8 Continuum concept

As noted earlier, classifications impose boundaries on continuous variations in channel planform. Leopold and Wolman (1957) recognized this explicitly, arguing that channel pattern reflects interactions among continuous variables such as sediment grain size and volume, gradient, boundary erodibility, and flow energy. They proposed that a continuum of channel patterns exists, with each pattern being defined by a combination of control variables.

Efforts to identify the most important control variables and consistent correlations with channel pattern include, but are not limited to

- slope versus discharge (Leopold and Wolman, 1957) (Figure 5.11), under the assumption that discharge is an independent variable and slope of alluvial channels reflects roughness, particle size, and drainage area;
- the percentage of silt and clay in the channel boundaries (Schumm, 1963), because the presence of cohesive sediment strongly influences channel stability, shape, and sinuosity;
- the ratios of depth to width (*d/w*) and slope to Froude number (*S/Fr*) (Parker, 1976), as a means of including channel form and flow energy parameters; and
- unit stream power versus median bed grain size (D_{50}) (Van den Berg, 1995), as two boundary conditions that are nearly independent of channel pattern.

Brotherton (1979) and subsequent investigators have argued for the importance of the relative ease of eroding the channel banks versus transporting bank material. Where transport dominates, the channel remains straight, whereas erodible banks facilitate braiding, with meandering as an intermediate scenario (Knighton, 1998).

In general, the progression from laterally stable (straight) channels through meandering to braided channels tends to correlate with increasing stream power, increasing *w/d* ratio (and hence increasing bank erodibility), and increasing volume and grain size of bedload (Ferguson, 1987). Although

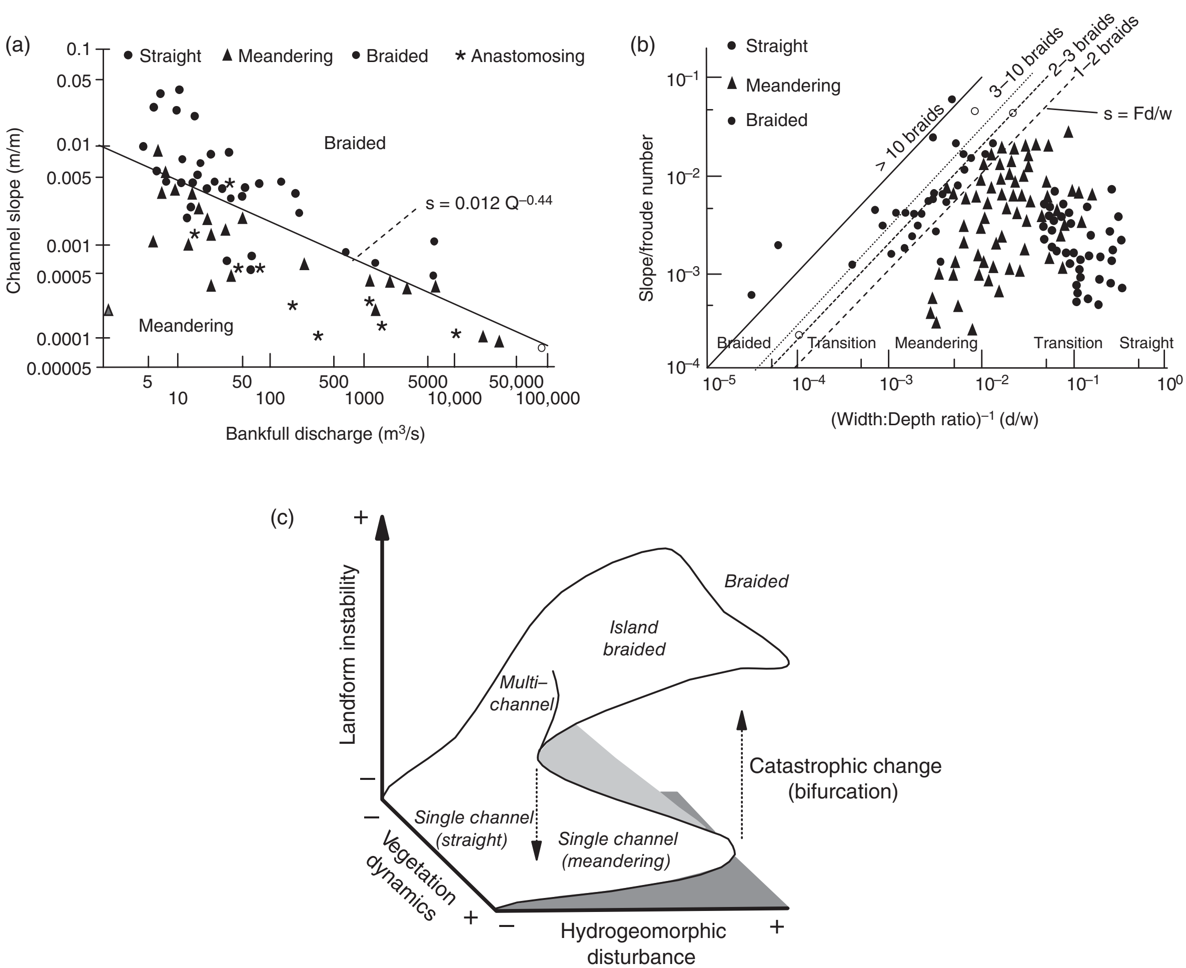


Figure 5.11 Different approaches to distinguishing channel planforms. (a) Braided and meandering with respect to slope versus discharge (from Leopold and Wolman, 1957). (b) Multiple channel types distinguished via slope/Froude number versus width/depth (from Parker, 1976). (c) Schematic representation of the continuity among different channel planforms, shown here in terms of a cusp catastrophe model (from Thom, 1975). This type of representation highlights how channels close to a threshold can have very different planforms (e.g., island braided vs. meandering), a condition in which a small change in the controlling variables, such as hydrogeomorphic disturbance or vegetation dynamics, can produce an abrupt change in channel planform (from Francis et al., 2009b, Figure 5).

plots such as Figure 5.11 over-simplify the complexity of channel planforms, these plots do provide useful insights regarding a given channel segment's proximity to a planform threshold and its likely response to relatively small changes in external variables (Schumm, 1985; Knighton, 1998).

Transitions between different planforms, and the ability for channels to repeatedly cross thresholds as conditions of boundary resistance and flow energy change, are particularly well represented using a cusp catastrophe model such as that in Figure 5.11c. By representing channel boundary resistance on the x and z axes, and magnitude, frequency, and duration of flow energy on the y axis, this model incorporates the balance between hydraulic driving forces and substrate resistance in a manner that facilitates understanding of the relative importance of individual variables influencing channel planform.

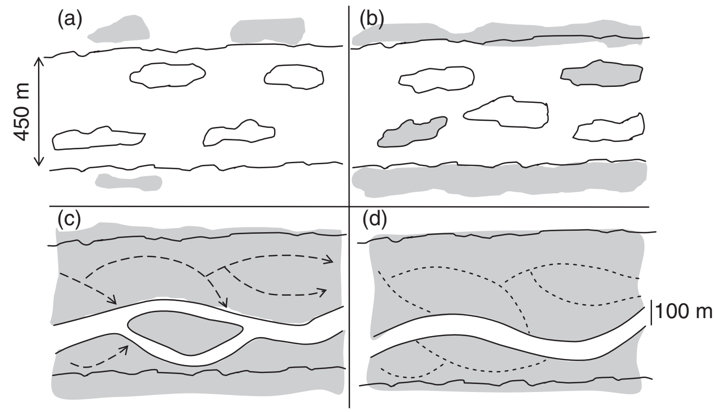


Figure 5.12 Schematic plan view illustration of river metamorphosis on the Great Plains of the United States. (a) Early 1800s: highly seasonal discharge maintains a broad, shallow braided channel with unvegetated bars. (b) Late 1800s: flow regulation creates more consistent flows with higher base flow and lower flood peaks, allowing riparian vegetation (gray shading) to establish along the channel banks and on some of the braid bars. (c) Early 1900s: droughts and flow regulation allow vegetation to establish below mean annual high water level, bars become islands, and single thalweg is dominant. Dashed lines indicate vestiges of historic channels that remain on floodplain. (d) Contemporary channel: bars and islands have become vegetated and attached to the floodplain, creating a wide, forested riparian corridor with a single narrow channel. (From Nadler and Schumm, 1981, Figure 10.)

5.2.9 River metamorphosis

The idea of continuity between individual channel patterns is also reflected in the recognition that pattern can change abruptly in space and time. Schumm (1969) used the phrase *river metamorphosis* to describe the “almost complete transformation of river morphology” in response to natural or human-induced changes. River metamorphosis typically refers to a rapid change in channel planform. Numerous examples have been described:

- Klimek (1987) described metamorphosis of a Polish river when native vegetation was cleared across much of the drainage basin and dramatic increases in sediment yield caused a meandering river to become braided.
- Flow regulation, along with introduction of exotic riparian vegetation that grows more densely along river banks, resulted in increased bank stability and sediment trapping that caused river metamorphosis along rivers in the southwestern United States. This metamorphosis took the form of narrowing, deepening, and alteration of braided channels to meandering channels (Birken and Cooper, 2006; Reynolds et al., 2012).
- Similarly, river metamorphosis occurred along rivers of the US Great Plains in which flow regulation caused a decrease in peak flows and an increase in base flows. Changes in flow regime allowed native riparian vegetation to grow more densely along the channel banks, and the combined effects of loss of peak flows and increased bank resistance caused braided rivers to narrow to a meandering planform (Williams, 1978b; Figure 5.12).
- A final example of a biotic external change that triggers river metamorphosis involves the presence or absence of beaver. As noted earlier, when present in a river, beaver build dams that promote channel–floodplain connectivity, overbank sedimentation and floodplain wetlands, and an anabranching planform. Beaver have been removed by trapping throughout much of their historic range. In parts of the western United States, beaver have also been outcompeted by large herbivores such as elk that have dramatically

increased in population density following twentieth century removal of predators. Elk eat the same riparian woody species that support beaver, and the combined loss of beaver dams and heavy elk grazing of riparian vegetation can result in metamorphosis within two to three decades of an anabranching channel to a single incised channel with unstable banks (Beschta and Ripple, 2012; Polvi and Wohl, 2012). Where the effects of elk are somehow reduced (e.g., riparian grazing exclosures or reintroduction of predators) and beaver return to a site, the channel can again metamorphose to multi-thread within a decade (John and Klein, 2004).

River metamorphosis is not so much a model of how rivers adjust to changing external controls, as the concept that pronounced channel change can occur very rapidly and in response to limited external change—such as loss of flood peaks or change in biota—which then triggers numerous, nonlinear responses in river form.

A specific aspect of changing channel planform is the presence, geometry, and erosional and depositional processes occurring where two segments of channelized flow meet at confluences. Confluences are inherent in multi-thread planforms such as braided or anabranching rivers, but can also be very important where tributary channels join single or multi-thread main channels, or where secondary floodplain channels rejoin the main channel.

5.3 Confluences

Confluences between tributary and larger channels are ubiquitous in drainage networks, and confluences between subchannels are widespread in braided and anabranching channels. Confluences have received increasing attention in recent years because these highly turbulent locations strongly influence processes as diverse as contaminant dispersal and navigation (Gaudet and Roy, 1995).

Average and maximum values of flow velocity typically increase at confluences because the cross-sectional area downstream from the confluence is commonly smaller than the sum of the areas of the contributing channels. The presence of a flow separation zone downstream from the junction can further reduce effective cross-sectional area. The higher velocity can be associated with a scour zone in the bed. The size and location of the scour zone reflect the junction angle at the confluence and the ratio of discharge in the confluent channels: dimensionless scour depth increases as the junction angle increases and as the ratio of the minor tributary discharge to the mainstem discharge increases, as noted earlier for braided channels. A shear layer develops at the margin of the flow separation zone because of the low velocity in this zone relative to the main flow. A shear layer with significant vorticity also develops near the middle of the channel where tributary and mainstem flows merge (Figure 5.13). Turbulence associated with this shear layer may significantly influence the bed scour zone. Strong secondary circulation and helical flow also characterize confluences. Velocity typically decreases beyond the confluence zone as flow merges into the single receiving channel.

Bars commonly form at confluences. The location and type of bar varies in relation to planform geometry. Symmetrical confluences with similar angles between each tributary and the receiving channel commonly have a mid-channel bar downstream from the confluence that reflects deposition of sediment eroded from the scour zone (Best, 1986). Asymmetrical confluences typically have bars associated with the zone of flow separation, in which low velocity and recirculation promote sediment deposition (Figure 5.13). Among the most well-studied examples are the sand bars that form downstream from tributary junctions along the Colorado River in Grand Canyon, USA.

The Colorado River in Grand Canyon exemplifies a river within a laterally confined bedrock canyon. Smaller, steeper tributaries entering the Colorado deposit bars and fans, commonly with episodic deposition during flash floods and debris flows on the tributaries. Coarse-grained tributary deposits create rapids on the Colorado, and constrict the mainstem, leading to development of a shear layer, strong secondary circulation, and

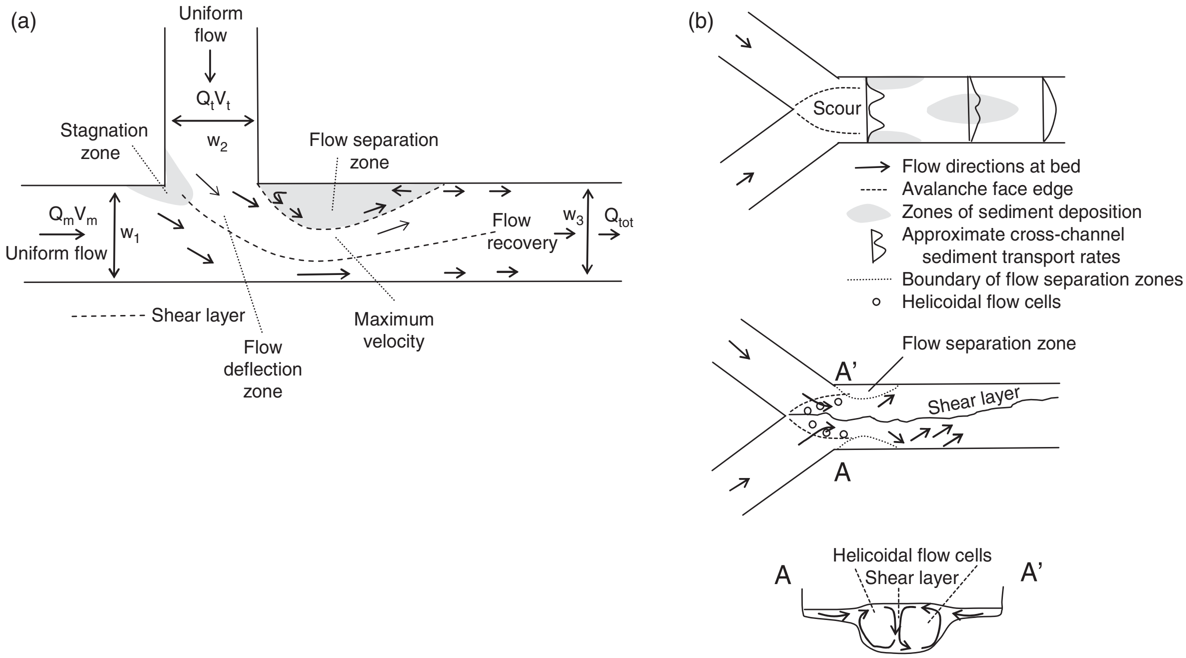


Figure 5.13 Hydraulics, cross-sectional geometry, and depositional features associated with channel confluences. (a) Conceptual model of flow dynamics at channel confluences, as derived from laboratory observations for an asymmetrical, 90 degrees confluence with main channel and tributary of equal width and depth. *Q* is discharge, *V* is average velocity, and *w* is channel width: Subscript *m* indicates the main channel and subscript *t* indicates the tributary. (From Best, 1987, Figure 4.) (b) Features of flow dynamics, bed morphology, and sediment transport at a symmetrical planform confluence. (From Mosley, 1976, Figure 3, and Best, 1987, Figure 5.)

finer-grained separation and reattachment deposits (Figure 5.14). The separation and reattachment deposits are typically sand bars that form important riparian habitat and recreational sites for thousands of people who float down the Colorado through Grand Canyon each year, and the eddy return-current channel creates a backwater that provides important habitat for endangered native fish (Schmidt and Graf, 1990). Flume experiments and hydraulic modeling indicate that the length of the separation zone reflects hydraulics and topography of the channel bed downstream from the confluence. Aggradation within the separation zone effectively decreases the length of this zone (Schmidt et al., 1993). The great majority of marginal deposition in a channel configuration such as that in Grand Canyon occurs in recirculation zones (Wiele et al., 1996).

Confluences between tributary and main channels in mountainous environments can also be strongly influenced by mass movements coming down the steeper tributaries. Low-order tributaries prone to debris flows can introduce abundant coarse sediment and wood into the main channel, resulting in more heterogeneous channel morphology and greater habitat diversity in the main channel (Benda et al., 2003a). Tributary inputs can have minimal effects where transport capacity of the mainstem is sufficient to rapidly redistribute tributary sediments. The likelihood that tributary inputs will significantly influence channel morphology on the mainstem increases with the size of the tributary relative to the mainstem (Benda et al., 2004)

Confluences can be characterized as *concordant* when two channels of equal depth join, or the more common *discordant confluences* when channels of different depth join (Robert, 2003). (Discordant confluences reflect differences in confluent channel dimensions as a result of differences in

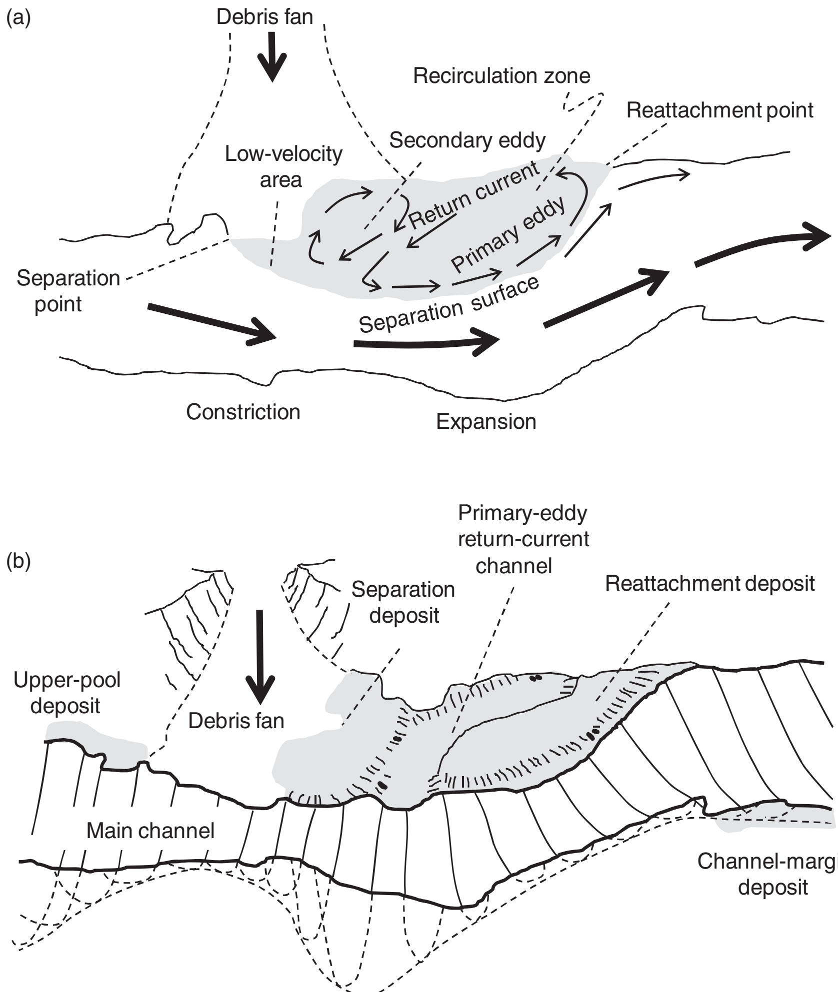


Figure 5.14 Separation and reattachment deposits in the Grand Canyon of the Colorado River. (a) Flow patterns. (b) Configuration of bed deposits. Gray shading indicates sand deposition. (From Schmidt and Graf, 1990, Figure 3, p. 5.)

discharge and boundary resistance; i.e., differences in hydraulic geometry.) The vertical mixing layer present in concordant confluences becomes more complicated at a discordant confluence, where the sudden drop in bed elevation from the tributary to the mainstem creates complex 3D flow and vertical upwelling (Best and Roy, 1991). The greater the bed discordance, the shorter the mixing length for the converging flows (Gaudet and Roy, 1995).

In regions with active tectonic uplift or relative base level fall, tributaries may not be able to incise as rapidly as the mainstem, leading to a pronounced drop at the tributary–mainstem confluence. In the most extreme case, this can result in a hanging valley (Section 5.4.3), with a substantial waterfall between the tributary and mainstem.

5.4 River gradient

The final level of adjustment in channel form occurs via changes in gradient. These can be changes in reach-scale gradient that occur over lengths tens to hundreds of times the average channel width. Gradient can also change over all or much of the longitudinal profile of a river.

Like channel planform, gradient can be largely imposed on a river by external constraints such as changing base level or erosionally resistant substrate, or gradient can be a response to existing water and sediment supply. In the latter case, gradient represents another dimension of channel adjustment and exhibits changes in response to varying water and sediment supply to the channel. The idea of gradient as a reflection of external inputs to the river is formally expressed in the concept of a graded stream.

The most widely cited definition of a *graded stream* is that proposed by Mackin (1948, p. 64): "A graded stream is one in which, over a period of years, slope is delicately adjusted to provide, with available discharge and the prevailing channel characteristics, just the velocity required for transportation of all of the load supplied from above." Mackin traced this idea back to Gilbert (1877) and Davis (1902a), although Davis viewed a graded stream as a condition developed only over very long periods of geologic time. Mackin, following Gilbert's perception, identified a graded stream as reflecting equilibrium, and applied this idea to the entire longitudinal profile of a river, as well as to more limited segments. The earlier definition of a graded stream implies that stream geometry is relatively constant over the time period of interest, although fluctuations about a consistent mean can occur.

A graded stream can develop relatively quickly—in some cases, over a matter of hours to a few days—in channels with readily erodible substrate. A graded condition typically requires longer to develop over an entire longitudinal profile or in more erosionally resistant substrate. Developing a graded condition over the length of a river requires that all portions of the river adjust to base level and that bed sediment along the river is distributed in adjustment with discharge. Before these adjustments can be accomplished along substantial portions of a river, however, boundary conditions (base level, sediment and water yield, substrate resistance) are likely to change. Developing a graded condition in more erosionally resistant substrate requires a sufficient frequency and duration of flows of high magnitude that exceed the threshold of boundary erodibility, and such flows are likely to be of longer recurrence interval. Rivers in the northwest Indian Himalaya, for example, display increased concavity downstream from glacially modified reaches and require more than 500,000 years to recover a graded condition (Hobley et al., 2010).

Discontinuities in process and form along a channel can occur in association with tributary junctions. These junctions can represent a substantial increase in discharge and/or an increase in the volume or average grain size of sediment (Section 5.6.1) supplied to the channel.

Discontinuities in channel process and form can also occur at transitions between bedrock, coarse-grained alluvial, and fine-grained alluvial substrates (Howard, 1980). The gradient of alluvial channels is commonly interpreted as reflecting hydraulic regime (Moshe et al., 2008). This condition is referred to as *transport limited* (Howard, 1994) because the transport of sediment is limited primarily by flow energy rather than sediment supply. In contrast, sediment transport in *supply-limited* channels is constrained primarily by the availability of sediment. Bed material grain size in a supply-limited channel adjusts by coarsening until the remaining exposed sediment is moved only in proportion to supply of that grain size. If sufficient supply is not available, bedrock is exposed along the channel, creating a subset of supply-limited conditions known as *detachment limited*. In detachment-limited channels, gradient may be less dependent on hydraulic regime because weathering must precede erosion (Howard, 1994, 1998).

Alluvial channel gradient shows a weak inverse correlation with discharge. Similarly, the relationship between gradient and median bed material size is not simple, but can be significant when sites of similar drainage area or discharge are compared (Hack, 1957). This indicates that gradient reflects both discharge and sediment supply (Knighton, 1998). This is logical, given that bed material size and discharge influence particle mobility, channel boundary roughness, rates of energy expenditure, and thus ability to adjust gradient. In channels with mixed grain sizes, questions arise as to which grain-size fraction exerts the strongest influence on ability to adjust gradient. Most investigators use some grain-size fraction coarser than D_{50} as better representing the influence of grain size on gradient in gravel-bed streams.

A threshold grain size of 10 mm separates sand-bed and gravel-bed streams (Howard, 1987). Gradient reflects quantity and grain size of load below this threshold, whereas gradient on channels with bed material coarser than 10 mm reflects threshold of motion of large grains rather than quantity of sediment. Sand-bed channels are referred to as *live-bed* or *regime channels* because sediment transport occurs at all but the lowest flows. Gravel-bed channels are known as *threshold* or *stable channels* because sediment moves only near bankfull discharge or during extreme flows (Howard, 1980). These channels commonly have gradients near the threshold of motion for the coarse grain fraction (Howard et al., 1994).

5.4.1 Longitudinal profile

Longitudinal profile typically refers to gradient along the entire length of a river from the channel head to the base level at which a river enters a larger river or a body of standing water. The concept of *base level* is particularly important in understanding adjustment of river longitudinal profile. First articulated by J.W. Powell (1875, 1876), base level can be conceptualized as a lever arm that influences channel bed elevation and gradient upstream from the base level. If base level increases in elevation, upstream gradient declines, transport energy declines, and sediment is typically deposited, causing the channel to aggrade to the new base level. If base level falls, gradient increases, and the channel incises to the new base level starting at the point of base level drop and propagating upstream.

Local base level can occur partway along a river, where the river enters a lake that has river outflow downstream. Local base level also refers to an erosionally resistant layer that limits upstream transmission of ultimate base level fall. Sea level forms the ultimate base level for rivers.

Relative base level change refers to a scenario in which the elevation difference between base level and a reference portion of a river changes. This can occur when tectonic uplift raises the drainage basin, for example, even if the ultimate base level of sea level remains constant during the period of uplift. Sea level has of course fluctuated dramatically during the Quaternary in association with the advance and retreat of continental ice sheets, causing ultimate and relative base level to change repeatedly.

In the description of longitudinal profiles, as in many other areas of fluvial geomorphology, G.K. Gilbert led the way with his three laws of land sculpture, which included the *law of divides*. This simply states that the gradient of a river steepens with proximity to the drainage divide (Gilbert, 1877). Although longitudinal profiles can be straight or convex (Figure 5.15), overall gradient of medium- to large-sized rivers typically decreases downstream, creating a concave longitudinal profile that Hack (1957) described as

$$S = kL^n \tag{5.7}$$

where S is gradient, k incorporates mean bed particle size, L is distance downstream from the drainage divide, and the exponent n is an index of profile concavity. S in this equation is a tangent to the curve that defines the relation between fall H and length L along a river (i.e., a longitudinal profile).

Based on field data from the eastern United States, Hack (1957) proposed an empirical form of Equation 5.7 using mean bed particle size (D_{50}, in mm) and distance downstream from the drainage divide (L, in miles)

$$S = 25\left(\frac{D_{50}^{0.6}}{L}\right) \tag{5.8}$$

For channel segments with constant D_{50}, integration of Equation 5.7 produces

$$H = k\, In\, L + c \tag{5.9}$$

where c and k are empirically derived constants.

(The integration is
$$S = kL^n$$
$$H = kL^n dL$$
$$= k\log_e L + c, \text{where } n \text{ equals } -1)$$

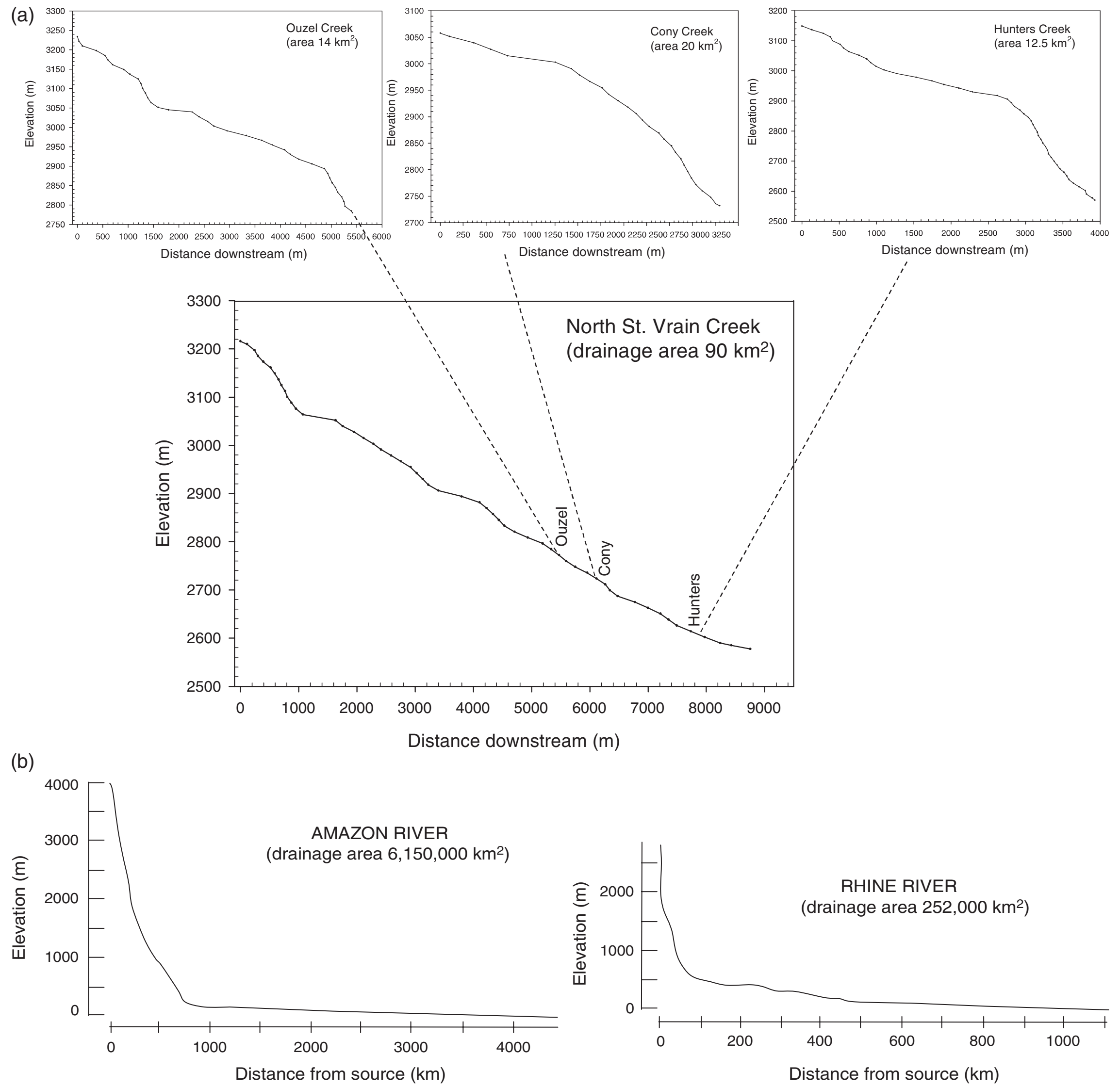


Figure 5.15 Sample longitudinal profiles for diverse rivers. (a) A mountainous river network in Colorado, USA. Ouzel, Cony, and Hunters Creeks are each tributary to North St. Vrain Creek at the points indicated on the longitudinal profile of North St. Vrain Creek. Each of these longitudinal profiles is relatively straight or convex, and the tributaries steepen as they enter the main valley of North St. Vrain Creek, which was glaciated during the Pleistocene. (b) The Amazon and Rhine Rivers, which display much more concave profiles. In each of these large rivers, the majority of elevation loss occurs in the mountainous headwaters.

In the much more common scenario in which particle size changes systematically downstream, (i.e., n does not equal -1), the equation takes the form of

$$H = \left(\frac{k}{n+1}\right) L^{(n+1)} + c \tag{5.10}$$

Snow and Slingerland (1987) used numerical simulations to demonstrate that the specific form of mathematical function that best approximates a river's longitudinal profile depends on the relative rates of change in water and sediment discharge and grain size downstream. Exponential, logarithmic, and power functions all create smooth, concave-upward longitudinal profiles, but the profiles of real rivers typically contain irregularities (Knighton, 1998).

The index of profile concavity (exponent n in Equation 5.7), as suggested by the correlations between reach-scale gradient and discharge and grain size, reflects rates of downstream change in discharge and substrate resistance. Profiles tend to be more concave where discharge increases rapidly downstream or grain size decreases rapidly. Increasing discharge translates to ability to transport the same bed-material load over progressively lower slopes. Similarly, decreasing grain size implies the ability to transport the available load over progressively lower slopes. Adjustments to increasing discharge or decreasing grain size can also occur in terms of bed roughness, cross-sectional geometry, and planform (Dust and Wohl, 2012b).

The longitudinal profile of a river can also be characterized in terms of the concavity index θ

$$S = k_s A^{-\theta} \qquad (5.11)$$

where S is channel gradient, k_s is the steepness index, and A is drainage area (Flint, 1974). This relation holds downstream of a critical drainage area. Critical drainage area varies, but is typically in the range of $0.2–0.9 \times 10^5$ m^2 (Whipple, 2004). Values of θ for bedrock channel segments vary from 0.3 to 1.2, and can be negative over short reaches. Concavity values <0.4 are associated with short, steep drainages strongly influenced by debris flows or with downstream increases in incision rate or rock strength that result in knickpoints. Moderate values of 0.4–0.7 correlate with active uplift and homogeneous substrate, and high values of 0.7–1 correlate with downstream decreases in rock uplift rate or rock strength (Whipple, 2004). The concavity index indicates the rate of decline in channel bed gradient with increasing drainage area: the higher the index, the faster the rate of decline in channel bed gradient.

As noted previously, even predominantly concave longitudinal profiles tend to exhibit irregularities in the form of local steepening. These irregularities can result from more resistant substrate, large point inputs of sediment, tectonic activity, or continuing response to base level fall. Although irregularities generated by any of these mechanisms can persist in alluvial channels, they are more likely to persist in bedrock channels. Alluvial channels respond more quickly than bedrock channels to local perturbations, and alluvial channels are more likely to adjust channel parameters such as cross-sectional geometry and planform, as well as gradient (Schumm et al., 1987).

Irregularities in the longitudinal profile have the potential to provide important insight into river history and adjustment. Consequently, they have been the subject of much attention, including how to quantify the irregularities using *SL* or *DS* indices.

5.4.2 Stream gradient index

A semi-log plot of the longitudinal profile of channel segments with a constant value of D_{50} should be a straight line, the slope of which (k) Hack (1957) referred to as the *stream gradient,* or *SL, index*. SL index can be calculated as the product of gradient (S) and total stream length (L) from the divide. This is the most common theoretical form of an equilibrium longitudinal profile against which actual profiles are assessed (Goldrick and Bishop, 2007).

Hack (1973) noted that abrupt spatial changes in stream gradient, as indicated by discontinuities in the longitudinal profile of a river, can be an equilibrium response to substrate variations, or a disequilibrium response caused by local deformation or upstream propagation of a knickpoint caused by relative base level fall. An equilibrium response to spatial variations in substrate erodibility should persist because steeper channel segments form in more resistant substrate. A disequilibrium response should be transient because reach-scale gradient decreases after the knickpoint migrates upstream. There is no inherent mechanism, however, for differentiating equilibrium steepening associated with greater substrate resistance from disequilibrium steepening associated with relative base level fall on an *SL* plot. Consequently, Goldrick and Bishop (2007) proposed the *DS* approach, derived from the power relationship between discharge and downstream distance, and the dependence of stream incision on stream power

$$H = H_o - k\left(\frac{L^{1-\lambda}}{1-\lambda}\right) \tag{5.12}$$

where H_o is an estimate of the theoretical elevation of the drainage divide if hydraulic processes were active right to the drainage head, k reflects factors such as equilibrium incision rate and stream hydraulic geometry as well as lithology, and λ is the exponent of the relationship between discharge and downstream distance. Equation 5.12 is essentially a reparameterized version of Equation 5.10, with $-k$ for k and $-\lambda$ for n. The *DS* form may be more appropriate for tectonically quiet areas than the *SL* form, because the *DS* form differentiates equilibrium steepening, which appears as a parallel shift in a *DS* plot, from disequilibrium steepening, which appears as disordered outliers on a *DS* plot (Goldrick and Bishop, 2007) (Figure S5.11).

5.4.3 Knickpoints

As noted previously, irregularities are more likely to persist in bedrock channels than in alluvial channels. Consequently, bedrock longitudinal profiles have been used as an index of rock uplift rate in *steady-state landscapes* that maintain statistically invariant topography and constant denudation rate (Whipple, 2001), and in landscapes experiencing transient increases in erosion rates (Snyder et al., 2000; Whittaker et al., 2008; Roberts and White, 2010). The assumption is that steeper portions of the profile reflect greater uplift. The ability of a river to maintain profile concavity in response to uplift, greater substrate erosional resistance, large sediment inputs, or base level fall can be a function of discharge, with larger rivers capable of responding more quickly and maintaining smoother, more concave profiles (Merritts and Vincent, 1989). Knickpoints, in particular, reflect an inability to maintain longitudinally continuous profile concavity.

A *knickpoint* is a step-like discontinuity in a river's longitudinal profile. A *knickzone* is a river segment steeper than upstream and downstream segments, but which does not have a pronounced vertical discontinuity such as a waterfall. Knickpoints can occur in weakly consolidated alluvium, but they are best developed in cohesive alluvium or bedrock. Knickpoints can be stepped, buttressed, or undercut, with headward erosion via parallel retreat or rotation of the knickpoint face (Figure 5.16). Knickpoints can migrate upstream from a site of base

(a) (b)

Figure 5.16 Persistent and ephemeral knickpoints along channels. (a) This bedrock knickpoint where a tributary enters the Wulik River in Alaska, USA, represents the inability of the tributary to incise through the bedrock as rapidly as the mainstem river (The Wulik River flows right to left in the foreground of this view). (b) This knickpoint along an ephemeral channel tributary to the South Fork Poudre River in Colorado, USA formed in response to enhanced water yield after a wildfire that burned the catchment 4 months before the photograph was taken.

level fall (Crosby and Whipple, 2006), or as a result of the increase in discharge relative to sediment supply. Knickpoints can form at a local base level created by more resistant substrate, in which case they are likely to disappear once the river incises an inner channel through the resistant material (Wohl et al., 1994). Knickpoints can also form where a large point source of sediment such as a landslide overwhelms fluvial transport capacity (Korup et al., 2006).

Rates of knickpoint retreat can be two orders of magnitude greater than erosion rates elsewhere along the channel (Seidl et al., 1997). Knickpoints are the geomorphic hot spots of incision along a river's longitudinal profile. Rates of knickpoint retreat can correlate with drainage area and thus follow the stream power law (Equation S4.1) (Bishop et al., 2005). Rates of retreat also reflect spatial variations in rock erodibility (Harbor et al., 2005) and sediment flux passing over the knickpoint lip (Lamb et al., 2007). Cosmogenic ^{10}Be dating of strath terraces downstream from headward-retreating knickpoints in western Scotland, for example, indicates that knickpoint retreat rates have declined since the mid-Holocene. Jansen et al. (2011) attributed this decline to a depletion of paraglacial sediment supply and a deficiency of abrasive tools to erode the bedrock knickpoints.

At the network scale, knickpoints at tributary junctions can create *hanging valleys* in which the tributary junction is perched well above the main valley bottom. Hanging tributary valleys are common in regions with alpine glaciation. Smaller tributary glaciers do not erode the valley as effectively as larger glaciers, leaving the tributary valley hanging when the glacial ice recedes (Figure 5.17). Hanging valleys can also form in tectonically active and incising landscapes with no history of glaciations. Under these conditions, rapid mainstem incision over-steepens tributary junctions beyond a threshold slope, or low tributary sediment flux during mainstem incision limits the tributary's ability to incise as rapidly as the mainstem (Crosby et al., 2007). The amount of over-steepening needed to form a hanging valley increases as tributary drainage area increases up to some maximum value, above

Figure 5.17 An example of a tributary hanging valley in a glaciated river network, here in Yosemite National Park, California, USA. The mainstem river flows left to right in the foreground.

which large tributaries can keep pace with base level fall and mainstem incision.

Crosby et al. (2007) predicted the maximum drainage area at which a tributary hanging valley, A_{temp}, can form

$$A_{temp} = \left(\frac{k_w k_q^b}{K_{GA} \beta} \frac{I_{max}}{U_{initial}} \right)^{\frac{1}{1-bc}} \qquad (5.13)$$

where k_w is a coefficient for the relation between channel width and discharge, k_q is a coefficient for the relation between discharge and drainage area, K_{GA} is a dimensional constant equal to (r/L_s), with r as the fraction of the volume detached off the bed

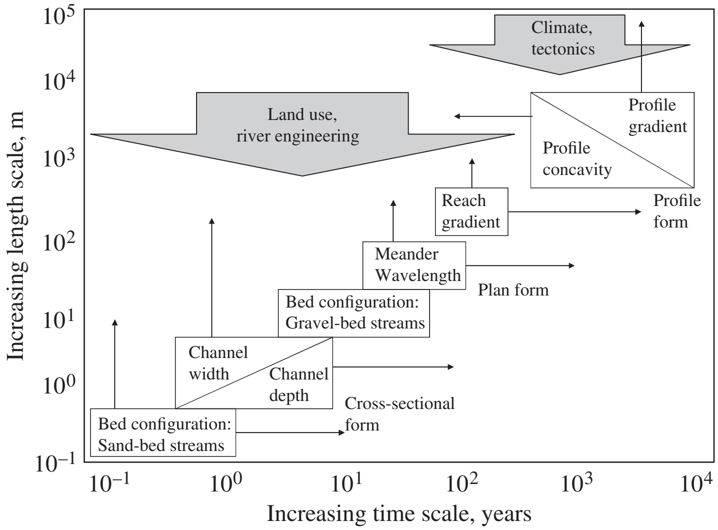


Figure 5.18 Temporal and spatial scales of river adjustment. (Modified from Knighton, 1998, fig 5.3, p 158).

with each collision and L_s as saltation hop length, β is the percentage of eroded material transported as bedload, I_{max} is maximum incision rate, and $U_{initial}$ is background rate of base level fall.

This equation describes a transient instability in a river with a profile that reflects incision dependent on sediment flux. Crosby et al. (2007) assumed that changes in sediment flux lag behind profile adjustment because the hillslope response that determines sediment flux depends on the transmission of base level fall through the network. Temporary hanging valleys form after base level fall if the mainstem transient incision rate exceeds the tributary maximum incision rate associated with the initial sediment flux. The terms in Equation 5.13 describe how the over-steepened reach A_{temp} adjusts as the main stem incision rate returns to just balancing the background rate of base level fall ($U_{initial}$) while the tributary continues to incise at the rate I_{max}, which is dependent on abrasion reflected in the variables of K_{GA} and β.

5.5 Adjustment of channel form

Feedbacks between channel processes and various aspects of channel geometry are implicit in much of the material covered thus far in this chapter and in preceding chapters. The magnitude and duration of a flow hydrograph influence the energy available to perform work in the channel and thus influence channel form, for example, but also respond to channel form as form influences travel time of flood pulses. Hydraulic forces influence sediment mobility and bedforms, but also respond to changes in boundary roughness caused by mobile sediment and bedforms. Rate of erosion governs whether a longitudinal profile can maintain concavity during tectonic uplift, but profile concavity also influences energy available for erosion.

Channel forms are typically described at the spatial scales of cross-sectional geometry, reach-scale planform and gradient, and basin-scale longitudinal profile. Adjustment of channel form can similarly occur at various spatial scales, including changes in *w/d* ratio or grain size and quantity of bed material, changes in sinuosity or number of subparallel channels (if the channel is not laterally confined), and changes in gradient and shape of the longitudinal profile.

Figure 5.18 illustrates the temporal and spatial scales over which various form components of rivers adjust to changing inputs and processes. This figure provides a very useful conceptual framework for thinking about feedbacks and adjustment in rivers because the diagram illustrates that

- smaller spatial-scale components of a river network (e.g., bed configuration) can change relatively quickly compared to larger scale components such as planform;
- for a given spatial scale or type of component, such as bedforms, the rate of adjustment increases as the erosional resistance of the material increases—in other words, live-bed channels and sand bedforms adjust more quickly than threshold channels and boulder-bed step–pool sequences;
- rivers include multiple components that can adjust to changes in boundary conditions (water and sediment input, substrate erodibility, base level), making it difficult to precisely predict the nature of river adjustment; and
- space and timescales over which adjustments occur in diverse river features overlap, implying that multiple aspects of a river are likely to respond to a change in boundary conditions, as implied by the expanded Lane's balance (Section 5.1.4).

Channel adjustment fundamentally represents some alteration of channel form in an attempt to reach equilibrium, as reflected in channel geometry and rate of flow energy expenditure, and as controlled by independent variables such as sediment and water yield, substrate resistance, and base level. The final two sections of this chapter examine (i) conceptual models developed to explain the physical processes that limit channel adjustment, and (ii) the high flows that are responsible for a substantial proportion of channel adjustment in many channels, respectively.

5.5.1 Extremal hypotheses of channel adjustment

Channel adjustment is sometimes conceptualized using *extremal hypotheses*, which are models based on the assumption that equilibrium channel morphology corresponds to the morphology that maximizes or minimizes the value of a specific parameter (Darby and Van De Wiel, 2003). Examples include minimization of rate of energy dissipation (Yang, 1976) or stream power (Chang, 1988), and the maximization of sediment transport rate (White et al., 1982) or friction factor (Davies and Sutherland, 1983; Abrahams et al., 1995). This approach stems from Langbein and Leopold's (1964) conceptualization of river form as the most probable state between the opposing tendencies of minimum total rate of work and uniform distribution of energy expenditure.

Extremal hypotheses have been criticized as being teleological and lacking explanatory power (Ferguson, 1986), but extremal hypotheses do explain a wide range of observations (Darby and Van De Wiel, 2003). The foundational assumption in extremal hypotheses is typically that a high rate of energy expenditure at a specified point in a channel will eventually result in boundary deformation until the rate of energy expenditure declines or the boundary is no longer adjustable by the energy available.

An example for steep, mobile-bed channels is the hypothesis that interactions between hydraulics and bed configuration prevent the Froude number from exceeding 1 for more than short distances or periods of time (Grant, 1997). Observed cyclical patterns of creation and destruction of bedforms (particularly antidunes) effectively maintain critical flow in these channels. Extremal hypotheses have also been used to explain the development of bank roughness in extremely deep, narrow bedrock slot canyons (Wohl et al., 1999), cross-sectional geometry (Eaton et al., 2004), and anabranching channel planform (Huang and Nanson, 2007), among other characteristics.

5.5.2 Geomorphic effects of floods

The geomorphic importance of floods of varying magnitude differs widely among rivers. Along rivers with relatively low hydrologic variability through time, the largest floods are unlikely to have persistent effects on channel and valley morphology because subsequent smaller flows quickly modify erosional and depositional features created during large floods. Large floods are more likely to create persistent effects as erosional resistance of channel boundaries increases and as hydrologic variability increases (Kochel, 1988), because subsequent smaller flows are less able to modify the geomorphic effects of large floods.

Within a region, large floods are typically more important in highland rivers than in lowland rivers (Froehlich and Starkel, 1987; Patton, 1988; Eaton et al., 2003). Highland rivers have steep gradients, narrow valley bottoms, coarse bedload, relatively flashy hydrographs, and less erodible channel boundaries, all of which magnify the geomorphic effects of floods (Kochel, 1988). Lowland rivers have larger buffering capacity associated with well-developed floodplains and greater drainage area, as well as finer bedload and channel boundaries that can be mobilized by smaller discharges. Floods can thus be of varying *geomorphic effectiveness*—defined as the ability of a flow to modify channel morphology (Wolman and Gerson, 1978)—across different segments of a river network.

The geomorphic effectiveness of a particular flood typically varies spatially along a river and between neighboring rivers in response to spatial variations in flood hydraulics, sediment supply, and erodibility of the channel boundaries (Miller, 1995; Cenderelli and Wohl, 2003; Procter et al., 2010). Antecedent conditions and the magnitude of a flood relative to earlier floods also influence geomorphic effectiveness (Eaton and Lapointe, 2001). The greater the difference in magnitude between a particular flood and previous flows, the more likely that particular flood is to create substantial channel change.

Erosional features created by large floods over cohesive substrates (Baker, 1988) include

- sculpted features (Section 4.3.3; Supplemental Section 4.1.3);
- inner channels (Section 4.1.3); and
- knickpoints (Section 5.4.3).

Flood erosional features in unconsolidated materials (Miller and Parkinson, 1993) include

- longitudinal grooves—elongate linear grooves parallel or subparallel to the local direction of flood flow, tens to hundreds of meters long, form in groups, individual grooves spaced 0.5–3 m apart, width and depth values range from centimeters to greater than a meter;
- channel widening and incision;
- stripped floodplains, which occur where general scouring that is not restricted to a well-defined scour mark or channel removes vegetation and fine-grained alluvium to depths of up to 1.5 m;
- anabranching erosion channels, which reflect incomplete channel widening that creates remnant islands in expanded channels;
- cutoff chutes in the form of well-defined channels that are typically several hundred meters long; and
- erosion of tributary fans impinging on the floodplain and main channel (Figure 5.19).

Depositional flood features include

- gravel bars within and along the margins of the channel;
- wake deposits in the lee of a large obstacle to flow;
- slackwater deposits of sediment deposited from suspension in areas of flow separation;
- terrace-like boulder berms; and
- aggradation within channels.

Depositional features in overbank areas include gravel splays and gravel and sand sheets (Miller and Parkinson, 1993). Gravel splays are lobate features in planform, with a convex profile. They are associated with severe channel or floodplain erosion, and are deposited where confined flow becomes unconfined, such as main-channel flow breaching a levee and entering the floodplain. Gravel and sand sheets are typically broader and thinner than splays. Floods can also drive substantial changes in channel planform, typically from single channel to braided (Scott and Gravlee, 1968; Cenderelli and Cluer, 1998).

The magnitude of channel change during a flood can be difficult to predict, but the longitudinal distribution of predominantly erosional and depositional flood features correlates strongly with

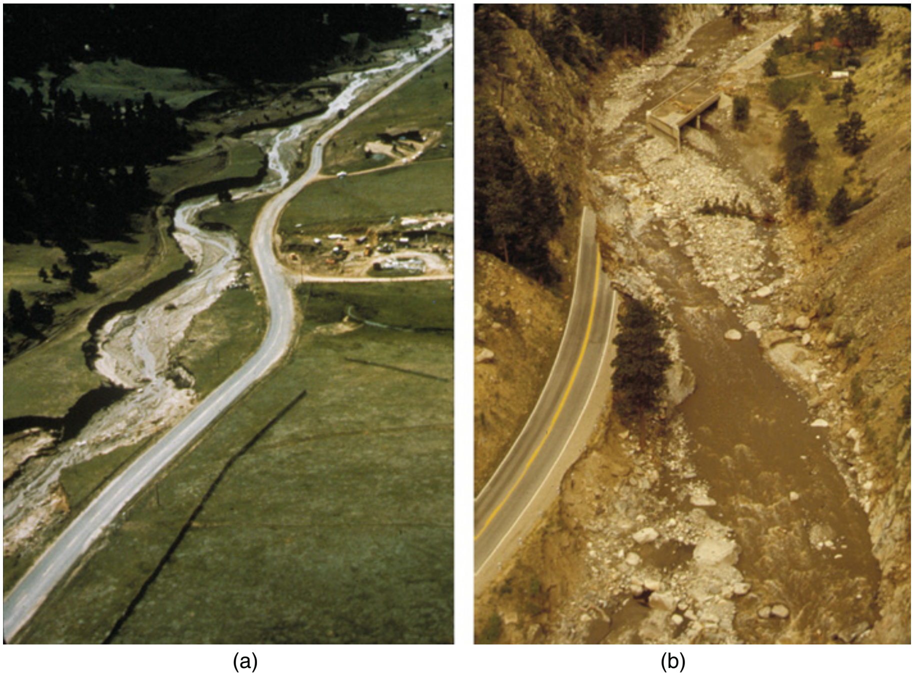


Figure 5.19 Examples of flood erosional features along the Big Thompson River in Colorado, USA during an extreme flood in 1976. (a) Channel widening along the upper, alluvial reaches of the river. (b) Channel widening in a bedrock-constrained canyon resulted in destruction of a bridge. Both photographs courtesy of Stanley A. Schumm.

valley and channel geometry and is thus predictable. Steep, narrow reaches are likely to experience, predominantly, erosion during floods, whereas relatively low-gradient, wide reaches become sites of deposition (Shroba et al., 1979). Despite the site-specific nature of erosion and deposition, various thresholds have been proposed for significant channel modification during floods, including a minimum stream power per unit area of 300 W/m^2 for low-gradient, alluvial channels (Magilligan, 1992) and an empirical power relation between stream power per unit area and drainage area ($\omega = 21\ A^{0.36}$) (Wohl et al., 2001). Costa and O'Connor (1995) proposed distinct thresholds for different types of substrates and channel modification during floods (Figure 5.20). If such thresholds can be quantified for a channel reach, and a measure of flow energy available to perform geomorphic work (e.g., excess shear stress or stream power) can be quantified over some time interval based on flood magnitude, duration, and frequency, then quantitatively estimating flood effectiveness should be possible.

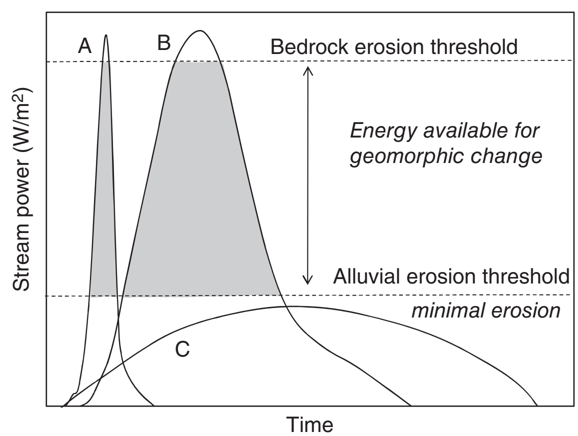


Figure 5.20 Schematic illustration of how the energy available for geomorphic change during a flood varies in relation to flood magnitude and duration. Hydrograph A represents a peaked, short duration flood of the type that results from a convective storm or damburst. Hydrograph B represents a longer duration, high-magnitude flood, such as might result from a cyclone. Hydrograph C represents a sustained, low-magnitude flood of the type that could occur during snowmelt. (From Costa and O'Connor, 1995, Figure 11, p. 54).

5.6 Downstream trends

Some components of channel form, such as cross-sectional geometry, typically exhibit consistent downstream trends as drainage area and, usually, discharge increase. Other components, such as bed-form configuration or bank stability, may or may not exhibit consistent downstream trends (Wohl, 2008b). The absence of consistent longitudinal trends typically results from spatial variation in other controls, such as tectonic regime, lithology, or climate, that override the effects of increasing discharge. Among those components of channel form that exhibit some downstream trends, but can also exhibit substantial longitudinal variability, are bed grain size and instream wood.

5.6.1 Grain size

Average bed grain size decreases downstream in most rivers at scales of tens to hundreds of kilometers—a pattern known as *downstream fining*. Downstream fining has been attributed to selective sorting (entrainment, transport, and deposition), abrasion in place or during transport, or some combination of these processes (Powell, 1998). The relative importance of selective sorting versus abrasion reflects factors such as clast erodibility. Abrasion is more important where clasts abrade readily (Parker, 1991).

In other settings, gravel-to-sand transitions occur over distances too short for substantial abrasion. Ferguson (2003) attributed these transitions to nonlinearities in bedload transport in which small increases in sand content lead to large increases in the mobility of sand and gravel. As sand deposition decreases the effective bed roughness, near-bed velocities increase, shear stress and turbulent kinetic energy decrease, burst and sweep events become less frequent, and the sand fraction remains more mobile than coarser particles. All of these factors contribute to creating a sand-bed channel downstream (Sambrook Smith and Nicholas, 2005), as do decreases in transport capacity (Ohmori, 1991; Parker et al., 2008).

Abrasion is commonly quantified using some form of Sternberg's (1875) law

$$D = D_0 e^{-\alpha x} \tag{5.14}$$

where initial grain size D_0 wears down to D at distance x from the origin at a rate given by the rock erodibility parameter α, which sets the abrasion length scale, $1/\alpha$ (Sklar et al., 2006). Values of α can vary by at least two orders of magnitude, from 10^{-3}/m to 10^{-5}/m, but tend to decrease as the bed material grows finer. In this general form of the Sternberg law, α incorporates both selective sorting and abrasion (Knighton, 1998). Fracturing of clasts is typically lumped under abrasion, although Chatanantavet et al. (2010) suggested distinguishing these processes (Figure S5.12).

Downstream fining can be interrupted by coarse sediment from hillslope or tributary inputs (Attal and Lavé, 2006) or from knickzone erosion (Deroanne and Petit, 1999). Mountainous catchments with close coupling of hillslopes and channels may have downstream grain-size distributions that closely mirror the size distribution of sediment supplied from hillslopes because local resupply from hillslopes offsets the influence of channel processes that create downstream fining (Sklar et al., 2006).

Mountain rivers can also have downstream coarsening. Grain size coarsens to a maximum value at the location along the channel where transport dominated by debris flows gives way to transport dominated by fluvial processes. Downstream fining occurs below this transition (Brummer and Montgomery, 2003).

Significant lateral sediment sources that influence downstream fining can be identified using drainage basin area, network magnitude, and the basin area–slope product to define individual channel links within which downstream fining occurs (Rice, 1998). The travel distance required to abrade coarser tributary sediments to the size of mainstem inputs from upstream can be predicted as

$$L^*_{\Delta D} = \frac{1}{\alpha} \ln\left(\frac{D_t}{D_m}\right) \tag{5.15}$$

where $L^*_{\Delta D}$ is the distance over which the grain size perturbation decays, D_t is mean grain size from the tributary, and D_m is mean grain size in the main-stem, and α is the rock erodibility parameter (Sklar et al., 2006).

Changes in cross-sectional geometry and transport capacity (Constantine et al., 2003; Rengers and Wohl, 2007) and river engineering (Surian, 2002) can also create substantial variation in downstream grain-size trends. Each of these studies in diverse environments indicates minimal downstream fining, or disruption or reversal of system-wide fining trends, along reaches in which the channel is laterally confined (and, typically, steeper).

In addition to decreases in average grain size downstream, bed material also becomes better sorted and individual particles become more rounded with distance from sediment source areas (Knighton, 1998). As with downstream fining, the distance over which these changes occur reflects clast erodibility and transport.

5.6.2 Instream wood

As noted in previous sections, wood within a channel has numerous geomorphic and ecological effects (Wohl, 2010; Gurnell, 2013) (Figure S5.13), including:

- increasing flow resistance and decreasing bank erodibility (Curran and Wohl, 2003; Manners et al., 2007);
- deflecting flow toward the channel boundaries and thus accentuating local boundary erosion (Hassan and Woodsmith, 2004);
- increasing channel width by deflecting flow toward the banks (Nakamura and Swanson, 1993);
- facilitating localized storage of sediment and organic matter by creating obstacles to flow and associated zones of flow separation (Andreoli et al., 2007);
- changing the dimensions of bedforms such as steps and pools—large instream wood pieces help to stabilize steps, for example, creating taller and more widely spaced steps than otherwise form in the absence of wood (MacFarlane and Wohl, 2003; Mao et al., 2008);
- creating forced alluvial reaches by creating backwater areas in which bedload is at least temporarily deposited—in the absence of the obstructions created by instream wood, the bedload would remain in transport (Montgomery et al., 2003b);
- enhancing overbank flooding and floodplain heterogeneity by creating in-channel obstructions that result in local aggradation and decreased flow conveyance (Jeffries et al., 2003);
- promoting bar growth and lateral channel movement (O'Connor et al., 2003); and
- initiating multi-thread morphology by enhancing magnitude and duration of overbank flows that can then erode secondary channels across the floodplain (Wohl, 2011c; Collins et al., 2012).

The magnitude and relative importance of these diverse effects vary downstream, as do the mechanisms of wood recruitment, transport, and storage (Gurnell, 2013). This can be illustrated using the relations proposed by Benda and Sias (2003) for wood dynamics within a channel segment of length x

$$\Delta S_c = \left[L_i - L_0 + \frac{Q_i}{\Delta x} - \frac{Q_0}{\Delta x} - D\right]\Delta t \qquad (5.16)$$

where ΔS_c is change in storage within the reach over time interval t; L_i is lateral wood recruitment into the channel; L_0 is loss of wood to overbank deposition during floods and abandonment of jams; Q_i is fluvial transport of wood into the reach and Q_0 is fluvial transport out of the reach; and D is *in situ* decay. Lateral wood recruitment reflects multiple forms of supply

$$L_i = I_m + I_f + I_{be} + I_s + I_e \qquad (5.17)$$

where I_m is chronic forest mortality, I_f is toppling of trees following mass mortality such as caused by fire or windstorms, I_{be} is inputs from bank erosion, I_s is wood from hillslope mass movements, and I_e is exhumation of buried wood from the floodplain (Benda and Sias, 2003) (Figure S5.14). Equation 5.16

is a wood budget, analogous to a sediment budget (Section 4.6).

In most regions, S_c decreases downstream. Wood load, typically quantified as volume of wood per unit area or unit length of channel, decreases downstream as transport capacity increases (Lienkaemper and Swanson, 1987; Bilby and Ward, 1989; Hassan et al., 2005). Simple ratios of wood piece length to average channel width and wood diameter to flow depth indicate relative transport capacity. Wood mobility increases as the average piece becomes shorter than channel width and as wood diameter becomes a progressively smaller fraction of flow depth (Lienkaemper and Swanson, 1987; Martin and Benda, 2001).

The relative importance of different mechanisms of lateral inputs also varies. Individual trees falling into the channel as a result of chronic forest mortality can be important in small channels, but become progressively less important as channel size increases. Bank erosion and floodplain exhumation become more important downstream as bank erodibility and the extent of floodplains increase. Hillslope mass movements add progressively less wood downstream as channels become buffered from hillslopes by wide valley bottoms and as topographic relief decreases. Lateral losses of wood also increase downstream as overbank processes and lateral channel movement grow more common.

Drag and buoyancy are the main forces mobilizing wood (Alonso, 2004). Wood entrainment is influenced by factors such as

- piece angle relative to flow direction—pieces parallel to flow tend to be more stable (Braudrick and Grant, 2000);
- presence of a root wad, which tends to decrease entrainment;
- wood density, which partly governs whether a log rolls, slides, or floats;
- piece length and diameter (Braudrick and Grant, 2000); and
- inclusion in a jam, which tends to stabilize individual pieces and decrease mobility (Wohl and Goode, 2008).

Most pieces move by floating, although waterlogged or otherwise particularly dense pieces can move in contact with the streambed. Transport distance increases with flow depth and duration above a floating threshold (Haga et al., 2002). Transport distance also increases as the spatial density of obstacles such as large boulders or stationary wood declines (Bocchiola et al., 2008). Wood rating curves that relate wood transport to discharge can be developed, analogous to sediment rating curves (Section 3.2.4), if records of wood transport can be developed using techniques such as video monitoring (MacVicar and Piégay, 2012).

Borrowing the concept of *alternative stable states* from ecology, Wohl and Beckman (2014), proposed that rivers in forested environments tend to be wood rich. Although recruitment of wood to the river fluctuates through time in response to disturbances such as forest fires, debris flows, or blowdowns, wood decay is sufficiently slow in most rivers of the temperate zone that at least some wood is always present in the stream. This wood helps to trap and retain newly recruited wood. If instream wood is removed or recruitment is eliminated for a period of time by clear-cutting, the river can enter a wood-poor condition in which the absence of instream wood reduces the likelihood that any subsequently recruited wood will be retained in the river rather than transported downstream.

Ecologists use alternative stable states to describe a scenario in which an ecosystem can exist under multiple states, or sets of unique physical and biological conditions. Alternative states are stable over ecologically relevant time spans, but ecosystems can transition from one stable state to another when perturbed sufficiently to cross a threshold. Collins et al. (2012) applied the concept of alternative stable states to floodplains along forested rivers.

Rivers are likely to transition from being transport limited with respect to wood in headwaters, to supply limited in larger channels (Marcus et al., 2002). Jams resulting from local decreases in wood transport increase downstream, whereas jams primarily forming around stationary wood pieces are more important in headwaters (Abbe and Montgomery, 2003; Wohl, 2013b) (Figure S5.15). The size and number of logjams are likely to be greatest in middle portions of river networks (Wohl and Jaeger, 2009). At the mid-basin transition from transport- to supply-limited conditions for wood, sufficient transport capacity exists to mobilize substantial volumes of wood, but local decreases in transport capacity associated with features such as mid-channel bars, zones of flow expansion, or channel bends (Braudrick and Grant, 2001) can trap large volumes of wood.

The geomorphic effects of individual wood pieces decrease downstream as each piece interacts with a progressively smaller proportion of the channel boundaries and cross-sectional flow, but aggregate effects of wood can be substantial in large channels. Historical accounts exist from diverse forested environments of enormous volumes of wood concentrated in portions of large alluvial rivers (Triska, 1984). These *log rafts* and *congested transport*—when numerous logs move together as a single mass, observed during flume studies (Braudrick et al., 1997)—significantly influenced channel and floodplain dynamics along many forested rivers prior to extensive modification by human activities such as timber harvest and snagging (removal of instream wood) (Wohl, 2013b). The volume of instream wood within most rivers flowing through forested drainage basins has decreased so substantially as a result of human activities that our perceptions of "natural" wood loads, and of the geomorphic importance of instream wood, are distorted (Chin et al., 2008).

5.7 Summary

Water and sediment are the drivers of channel form. We study hydraulics and sediment dynamics to understand channel form—to predict the channel form likely to result from specific engineering manipulations such as altered flow regime downstream from a dam or stabilization of banks through an urban area, and to interpret the environment that has resulted in an existing natural channel form. The fundamental questions in understanding channel form through time and space are (1) What magnitude and frequency of flow are most important in creating and maintaining channel form? and, (2) Through what processes, and how frequently, do changes in channel form occur? The relations between water, sediment, and channel form are not a simple, one-way process. Channel form responds to discharges of water and sediment, but form also influences the distribution of hydraulic variables and the dynamics of sediment in the channel.

Channel forms—from cross-sectional geometry to longitudinal profiles—represent the intersection of physics and history. The physical balance between hydraulic driving forces and substrate erodibility governs the creation and maintenance of river form. But the history of past channel adjustments and of larger factors as diverse as tectonics, climate, and land use, which influence driving forces and resistance, constrain the manner in which the balance is expressed. A river is a physical system with a history.

Chapter 6
Extra-channel environments

Extra-channel environments refer to fluvial erosional and depositional processes and the resulting fluvial landforms created or now found outside of active channels. These processes and landforms are contrasted with those within the boundaries of the active channel (Chapters 3, 4, and 5). This is an arbitrary distinction because of the close coupling between active channels and marginal areas of fluvial erosion and deposition, but the distinction serves to emphasize some of the unique aspects of floodplains, terraces, alluvial fans, deltas, and estuaries. All of these features can be laterally and longitudinally extensive and persistent landforms. As terrestrial–aquatic interfaces, they can also be chemically complex and biologically diverse and productive environments that are strongly influenced by biotic activities. Globally, these riverine environments are also heavily altered by human activities, with consequences for hazards and for ecological sustainability.

6.1 Floodplains

Floodplains are low-relief sedimentary surfaces adjacent to the active channel that are constructed by fluvial processes and inundated frequently. Floodplain relief is typically ~0.1–0.5 times the bankfull depth of the river (Dunne and Aalto, 2013). Engineers designate floodplains based on average recurrence interval of flooding, as in the 10-year floodplain or 100-year floodplain. Nanson and Croke (1992) refer to such surfaces as the *hydraulic floodplain*, because the designation does not imply anything about geomorphic history or fluvial influence. Geomorphologists historically referred to floodplains as those surfaces that are flooded at least once every 2 years, with the assumption that such surfaces are composed largely of fluvial sediments deposited under the current flow regime, rather than relict sediments deposited under very different conditions. Nanson and Croke (1992) describe this as a *genetic floodplain*. Even a genetic floodplain can be flooded less frequently than approximately once every 2 years, however, depending on the hydroclimatic regime. Analyzing data from 28 gaged sites on rivers in the western United States, Williams (1978a) found that the most frequent recurrence interval for floodplain inundation was about 1.5 years, but recurrence intervals varied from 1 to 32 years. Floodplains can form along river courses in valley bottoms, on alluvial fans, and on deltas (Bridge, 2003).

Floodplains can be examined in many contexts. They exert an important influence on lateral and longitudinal connectivity within a drainage network, buffering hillslope inputs, and commonly slowing the downstream movement of water, sediment, nutrients, contaminants, and organisms.

Rivers in the Landscape: Science and Management, First Edition. Ellen Wohl.

Companion Website: www.wiley.com/go/wohl/riverslandscape

Floodplains thus serve as reservoirs, in the sense of at least temporarily storing diverse materials. Storage time can range from hours to months for flood waters, and up to thousands of years for sediment.

Floodplains act as a safety valve during large floods. As flood water spills over the channel banks and across the floodplain, the water typically encounters greater flow resistance and moves more slowly, facilitating sediment deposition. Slower overbank flow attenuates the flood, creating a peak discharge of lower magnitude and longer duration. Turbulence and flow separation between the overbank and channelized flow dissipate energy and reduce channel-bed and bank erosion.

Floodplains typically enhance the storage of sediment and solutes over diverse time spans that depend on factors such as frequency of overbank flow and width of the floodplain relative to rate of lateral channel migration. Sediment and solutes entering the floodplain can come from upland environments, upstream or tributary portions of the river network or, particularly in the case of solutes, from groundwater sources.

Floodplains also tend to be biologically rich habitats. Many aquatic organisms use the floodplain during periods of inundation for breeding, nursery, and feeding areas. The flush of nutrients that returns to the channel as flood waters recede helps to support aquatic communities in the channel (see Section 3.2.4). The diversity of floodplain habitat and moisture levels commonly supports more diverse plant communities than are found in the adjacent uplands.

The water inundating a floodplain does not necessarily come exclusively from overbank flow (Figure S6.1). Flooding can occur before natural levees overtop because flood waters can enter a floodplain from rising groundwater, overland flow from adjacent slopes, and tributaries, as well as overbank flow from the main channel. Mertes (2000) distinguished inundation patterns on dry floodplains, where most water overflows channel banks, from saturated floodplains in which rising groundwater can contribute significantly. Flooding associated with rising groundwater is evidenced by increasing depth of standing water before the flood wave arrives, and by clear water that appears distinctly different from the turbid flood waters of the main channel.

An example of the influence of non-river water on flooding comes from the Altamaha River of Georgia, USA, where measurements during several floods revealed that the floodplain was inundated by water carrying virtually no sediment while the river was still rising but below bankfull level (Mertes, 1997). Deep ponding on the floodplain prior to overbank flow limited the ability of overbank flow to spread laterally across the floodplain once the river overtopped its banks. On this sinuous river, overbank flow entered the floodplain at the inflection point of river bends, flowed obliquely onto the floodplain, and retained a significant down-valley component. The river water moved down the floodplain in a relatively straight line between river bends and covered only a narrow zone (~2–4 km) on a floodplain up to 10 km wide (Mertes, 1997).

The *perirheic zone* refers to the zone of mixing surrounding the flowing river water (Mertes, 2000). Mixing between flooding river water and locally derived water is governed by characteristics such as the spatial density of floodplain channels: high density limits mixing because the floodplain channels contain flood waters leaving the main channel.

The Amazon River in some respects exemplifies floodplain dynamics. As the largest river in the world, the Amazon has enormous floodplains. As a largely unregulated river without human-built levees, the Amazon still experiences a natural flooding regime and high channel–floodplain connectivity. Limited mixing occurs in the upper reaches of the Amazon River, which have a high density of floodplain channels, whereas the middle reaches have enhanced mixing from overbank flows across levees along the main channel. The lateral extent of the incursion of river water into the floodplain varies from about 5 km in upstream reaches of the Amazon to 5–15 km in middle reaches. Mixing is also limited in the lower reaches of the Amazon, where large floodplain lakes have sufficient hydraulic head to prevent river water from entering large portions of the floodplain (Mertes, 2000).

Floodplains along large, lowland rivers such as the Amazon can be extremely extensive (Figure S6.2). Floodplains extend for approximately 4000 km along the Amazon or the Nile River, and can extend up to 100 km across. Dunne and Aalto (2013) distinguished the modern, most recently active *channel belt landforms* on a floodplain from the wider floodplain that can be of Holocene or even Pleistocene age. The channel belt along a meandering or braided river can be up to several times the width of the channel bends, and such bends can be up to 100 times the bankfull channel width. Channel belts on multi-thread large rivers can be even wider.

Active floodplains were historically much more extensive. People have systematically reduced the extent and duration of overbank flow by

- regulating the flow and removing flood peaks to the extent possible;
- constructing levees or transportation corridors (Blanton and Marcus, 2009) that physically block access to the floodplain; and
- enlarging channels so that higher flows remain within the channel.

Floodplains have been particularly targeted by these activities because extensive and prolonged overbank flooding is regarded as a nuisance or a hazard that limits human access to the flooded areas or travel across these areas, and because floodplains have long been regarded as highly desirable locations for human communities as a result of fertile soils and easy access to water supply and water-based transport. Many of the world's great cities are largely or wholly within floodplains, including London (capital of the United Kingdom), Seoul (capital of South Korea), Paris (capital of France), Tokyo (capital of Japan), Dhaka (capital of Bangladesh), Caracas (capital of Venezuela), and Kinshasa (capital of the Democratic Republic of Congo). Obviously, governments in every continent think they can control flooding sufficiently to continue to base themselves within a floodplain. The location of major cities within floodplains can result in tremendous damage and loss of life when flow regulation, levees, or channelization fail for some reason. In addition to the obvious hazard of flooding, lateral channel migration and avulsion can create substantial change, creating hazards for people and structures.

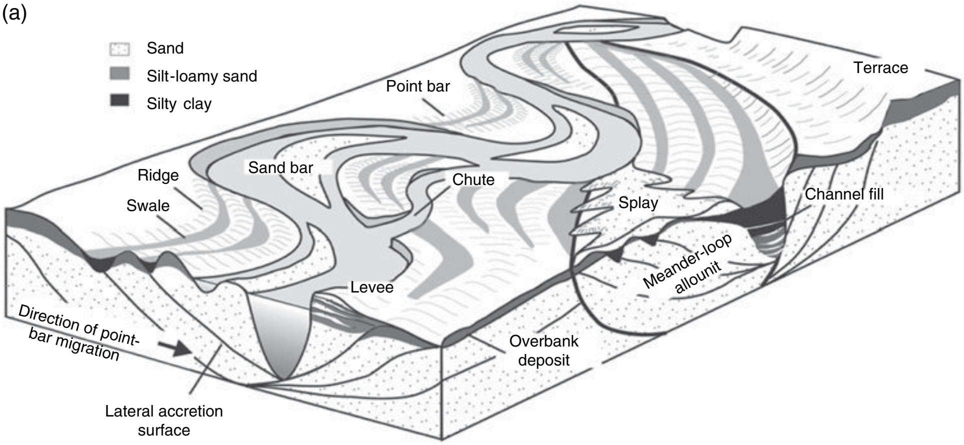
Relatively few large rivers with largely unaltered flood regimes and floodplains remain. Among these are the Amazon and the Congo, the two largest rivers in the world, and the great northern rivers the Lena and Mackenzie, which are rated as moderately affected by dams, as well as the Yukon, which is unaffected (Nilsson et al., 2005). The commonality among these rivers is that they all have low average population density within the drainage basin. In contrast, large rivers such as the Mississippi, Danube, Nile, Murray–Darling, and Yenisey, which historically had large floods and extensive floodplains, have lost flood peaks as a result of flow regulation, and lost floodplains to disconnection from the river and human settlement and land use within the floodplain.

6.1.1 Depositional processes and floodplain stratigraphy

Floodplain form reflects a history of both erosion and deposition, but the majority of floodplains are predominantly depositional environments that store large volumes of sediment for varying lengths of time. Floodplain deposits can be categorized as *vertical accretion* deposits that form when sediment settles from suspension in the lower-energy environment of the floodplain or as lateral accretion deposits. Vertical accretion deposits include levee and backswamp or flood basin sediments.

Sedimentation rates typically decrease across the floodplain with distance from the channel (Swanson et al., 2008), and active sedimentation extends no farther than a few kilometers from the channel even on very large floodplains (Dunne and Aalto, 2013). Rates of vertical accretion reported for diverse rivers range from 0 to nearly 6000 mm per year (Knighton, 1998), but are typically at most a few centimeters per year. Sedimentation is commonly extremely slow at distances of more than a kilometer from the main channel, and these lower elevation portions of

(a)



(b)

Figure 6.1 (a) Idealized diagram of a large, sinuous river floodplain, showing types of lateral and vertical accretion deposits discussed in text. (From Holbrook et al., 2006, Figure 2, and Dunne and Aalto, 2013, Figure 21.) (b) Levee formed along a distributary channel of the Slave River in northern Canada. The Slave River forms an extensive delta as it enters Great Slave Lake and this photograph was taken several kilometers upstream from the distributary channel's entry into the lake. Arrow indicates flow direction in channel and dashed line indicates approximate shape of levee, which has been truncated in the foreground in order to create a parking area and boat access to the channel.

the floodplain can be covered by lakes or swamps (Dunne and Aalto, 2013). In regions wet enough to support at least partial vegetation cover, floodplain deposits commonly include organic matter transported by flood waters and varying in size from fine particulate organic matter (~0.5–1 mm in diameter) to large tree trunks. Floodplain sediments are also typically bioturbated by plant roots and animals (Bridge, 2003).

Natural *levees,* as distinct from levees constructed by humans, are discontinuous, linear features immediately adjacent and parallel to the channel banks that form where the coarsest suspended sediment is deposited as velocity drops in the mixing zone between channelized and overbank flows (Figure 6.1). Sediment accumulates faster along the levee than it accumulates as flow spreads farther from the channel. This results in a continuous berm that gradually declines in elevation with distance from the channel (Dunne and Aalto, 2013) and can be up to four channel widths across (Bridge, 2003). Levee height commonly decreases downstream as the grain size of sediments in suspension decreases (Dunne and Aalto, 2013).

Backswamp or *flood basin* deposits are typically finer-grained sediment—wash load—settling from suspension across the portions of the floodplain between the levees and the valley walls or the outer limit of flooding. Flood basins are typically much longer than wide, and segmented by alluvial ridges and crevasse splays (Bridge, 2003). Sediment deposited in this environment can fill abandoned or secondary channels (known as sloughs, billabongs, and other names in different parts of the world), or can be deposited more uniformly across the floodplain flats. Very large depressions on the flood basin can contain lacustrine (lake) deposits and deltas. Ephemeral lakes in arid regions can contain evaporite minerals. Backswamp deposits are typically rich in organic matter, which can be autochthonous (derived from plants growing at the site) or allochthonous (transported from upland or upstream sites and deposited in the backswamp environment). Flood basins form the lower elevation areas of floodplains.

Alluvial ridges are slightly higher areas formed of active and abandoned channels and bars. The ridges include accretionary features, levees, and crevasse channels and splays (Bridge, 2003).

Splay or *crevasse splay deposits* occur when a levee is breached by piping or overtopping. The breaching process creates a narrow cut through the levee that contains higher velocity flow able to transport coarser suspended sediment into the backswamp area. Sand splays formed in this manner can be quite extensive. The 1993 flood on the Mississippi River created sand splay deposits greater than 60 cm thick that extended hundreds of meters in length and breadth (Jacobson and Oberg, 1997).

The other major category of floodplain deposits is associated with lateral accretion. *Lateral accretion* deposits result from sediment deposition along the margins of the active channel: this sediment can become part of the floodplain as the channel migrates laterally. Lateral accretion deposits include islands and bars, as well as lag deposits of the channel-bed sediment. The classic example is point bar sediment incorporated into the floodplain as a meander bend migrates toward the outer, cut bank of the bend. Lateral accretion deposits can form the bulk of floodplain sediment in gravel-bed rivers that are laterally mobile.

Well-developed bars can concentrate flow along the outer margins of bends and accelerate bank erosion and bend growth (Legleiter et al., 2011). Bend migration results in lower shear stress and sediment transport across the bar, which facilitates growth of vegetation and accumulation of finer sediment from suspension, raising the bar surface and blocking minor channels that might have developed across the bar (Braudrick et al., 2009). As bars are incorporated into the floodplain, sequences of arcuate bars with intervening, poorly drained depressions create *scroll-bar topography* that can persist for centuries as differences in soil grain size and moisture support different plant communities (Figure S6.3).

The lateral accretion of bars in a braided river creates a *braid plain* of channel-margin sediments. This plain, which is analogous to the channel belt of a sinuous river, has an irregular surface and thin, discontinuous, vertically accreted sedimentary covers in depressions and former channels (Dunne and Aalto, 2013).

The juxtaposition of different types of deposits creates vertical relief on floodplains. A floodplain created primarily by lateral accretion is sometimes known as a *flat floodplain* (Butzer, 1976). A flat floodplain consists mainly of bed-material deposits—either bedload in gravel-bed rivers, or suspendible sandy or silty bed material in large, lowland rivers (Dunne and Aalto, 2013). This type of floodplain contrasts with what Butzer (1976) called a *convex floodplain* along rivers dominated by vertical accretion. In a convex floodplain, the thickest and coarsest deposits close to the channel create a slightly higher floodplain elevation adjacent to the channel than at the margins of the floodplain.

Floodplains can also include non-fluvial sediments intermixed with the fluvial material. Eolian sediments can be interbedded with fluvial floodplain deposits in arid regions as a result of eolian input from adjacent regions, or reworking of unvegetated floodplain sediments. Colluvial sediments can be deposited on the outer margins of a floodplain with steep valley walls.

Floodplains thus form from a combination of within-channel and overbank deposits, and the relative importance of these two basic depositional environments varies in relation to factors such as rate of lateral channel movement and frequency, extent, duration, and sediment concentration of overbank flows. In an influential paper on floodplain deposition, Wolman and Leopold (1957) proposed that lateral accretion deposits dominated most floodplains, accounting for up to 90% of the total sediment deposited. Subsequent research, however, suggests that the situation is more complicated and varies greatly among different rivers.

The timescales of vertical and lateral floodplain accretion can differ, and rapid lateral channel migration can remove gradually accumulated vertical accretion deposits. The relative importance of lateral and vertical accretion can alter through time in response to

- changes in the volume and grain size of sediment available during floods—finer-grained sediment carried in suspension is more likely associated with vertical accretion, and coarser material moving in contact with the bed is more likely associated with lateral accretion;
- transport energy of the flood waters—higher transport energy can result in overbank velocities too high to permit settling from suspension and vertical accretion; and
- geometry of depositional sites—floodplains with numerous depressions that enhance locally low velocity and settling from suspension can promote vertical accretion.

The relative importance of vertical and lateral accretion can also vary downstream. Along the Brazilian portion of the Amazon River, floodplain deposition in the upstream reaches results from lateral accretion in floodplain channels. Overbank deposition dominates downstream portions of the floodplain (Mertes et al., 1996).

As a relatively well-studied and still-natural river, the Amazon River provides insight into some of the aspects of floodplain sedimentation characteristic of very large rivers. The huge drainage area and long translation times for passage of a flood wave along large rivers create gradually rising and falling hydrographs. Overbank flow can last from many weeks to as long as 6 months (Richey et al., 1989). The combination of long duration overbank flows and abundant fine-grained particles with low settling velocity can result in large transfers of sediment into the floodplain (Dunne and Aalto, 2013).

Recent work emphasizes biotic influences on floodplain form and process. Riparian vegetation growing on bars can help to stabilize and enlarge the bars, facilitating lateral accretion to the floodplain (Gurnell and Petts, 2006). Formation of channel-spanning obstructions such as beaver dams and logjams can lead to local aggradation and loss of conveyance that enhances overbank flow. This can in turn lead to the formation of multi-thread channels, as well as enhancing vertical accretion (Collins and Montgomery, 2002; Sear et al., 2010; Westbrook et al., 2011; Wohl, 2011c; Polvi and Wohl, 2012). Collins et al. (2012) conceptualized these influences in terms of a *floodplain large-wood cycle* in which logjams create alluvial patches and protect them from erosion, providing sites for trees to mature over hundreds of years in valley bottoms where floodplain turnover is typically much faster. These patches of old-growth forest in turn provide large wood to rivers that can form persistent logjams, thus perpetuating the cycle. Floodplains influenced by this process have multi-thread channels separated by surfaces of different age and elevation, as well as a mosaic of forest stands of varying age (Figure 6.2).

Episodically forming channel-spanning logjams can also facilitate greater depth and duration of overbank flow and floodplain aggradation. Oswald and Wohl (2008) described how jökulhlaups (outburst floods) from a headwater glacier in the Wind River Mountains of Wyoming, USA, create peak flows much greater than the typical snowmelt peak flow. Higher peak flows enhance bank erosion and wood recruitment, creating numerous logjams. Jökulhlaups occurring every few decades create logjams that break up over a period of a decade, but the overbank aggradation that occurs while the logjams are present can persist over time periods of 10^2–10^3 years (Figure S6.4).

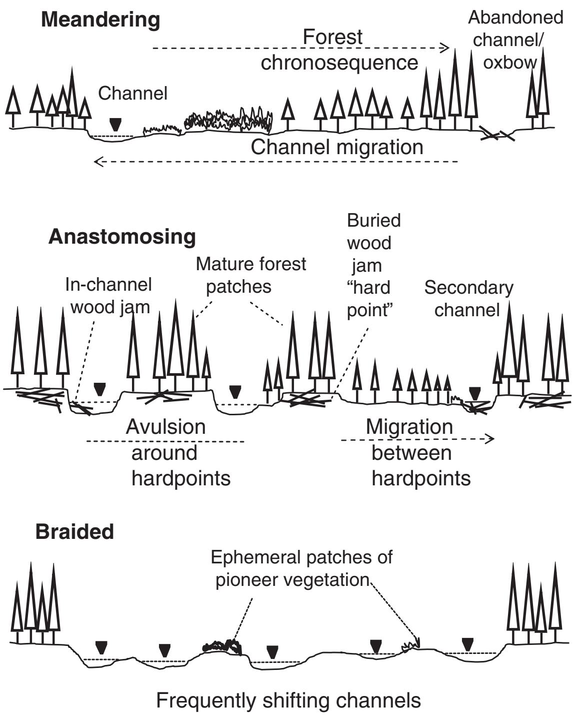


Figure 6.2 Cross-sectional views of three idealized models of floodplain landforms and forest development for river valleys of moderate gradient in the US Pacific Northwest. The meandering river model is dominated by lateral migration of meanders and associated meander cutoffs. These processes create scroll-bar topography, oxbow ponds, and sloughs. Plant germination on newly accreted sediments, and continued lateral migration and forest succession, create chronosequences of surface and forest ages. Maximum forest age is limited by the rate at which the river meanders across the floodplain. In the anastomosing river model, stable alluvial deposits associated with wood jams can resist lateral channel erosion for hundreds of years, providing sites for tree germination. Maximum forest age is limited by the stability of these deposits. Channels avulse around stable patches and migrate as in the meandering river model, creating a mosaic of multiple channels and floodplain elevations, and greater diversity in forest patch age. Braided rivers are dominated by multiple, frequently shifting channels. Mature forest vegetation is limited to the channel margins, although ephemeral patches of pioneer vegetation can grow on braid bars within the channel network. (From Collins et al., 2012, Figure 3.)

Although the conceptual model of a floodplain large-wood cycle was developed specifically for moderate-sized rivers in the Pacific Northwest region of the United States, it is important to remember that forested floodplains were once much more common. In his classic 1837 textbook, *Principles of Geology*, Charles Lyell commented on the enormous *log rafts* characteristic of large American rivers in the lower Mississippi and Atchafalaya of the southeastern United States. Historical descriptions indicate that these log rafts occurred on rivers across the United States, including the northeastern conifer forests, the temperate deciduous forests of the north-central region, and the humid temperate rainforest of the Pacific Northwest (Wohl, 2013b). Similar log rafts may have occurred historically on continents besides North America.

Lyell specifically mentioned how log rafts in the southeastern United States, which could stretch for >200 km along a river (Phillips and Park, 2009), enhanced overbank flooding. A sedimentary facies suite attributed to low-energy, organic-rich rivers with multiple, anabranching or anastomosing channels and stable alluvial islands first appeared during the Carboniferous, when tree-like plants could create channel blockage by logjams (Davies and Gibling, 2011). In other words, instream wood has been influencing flow resistance, channel conveyance, and overbank flows for a very long time, and these influences extended across the forested regions of the world. Wohl (2013b) proposed that widespread removal of instream wood during the past two centuries resulted in a fundamental change in floodplain dynamics because of the associated decrease in overbank flows and sedimentation, as well as decreased avulsion and formation of multi-thread planform (Figure S6.5).

Beavers were also once extremely abundant and ubiquitous in forested environments of North America (*Castor canadensis*) and Europe (*Castor fiber*). The channel-spanning dams that these animals built on smaller rivers certainly influenced magnitude and duration of overbank flows, and thus floodplain dynamics.

The balance through time among vertical accretion, lateral accretion, and lateral channel migration is reflected in floodplain stratigraphy. Sinuous channels typically create a characteristic *fining-upward sequence* (Figure S6.6). As a meander bend migrates laterally, a geographic location that starts as a deeper portion of the channel near the outside of the bend transitions to a point bar environment and eventually to a floodplain. The stratigraphic sequence at that location thus fines upward from relatively coarse channel-lag deposits, to slightly finer-grained point bar deposits, which are in turn capped by even finer overbank deposits. Floodplains created by sinuous channels also commonly have predictable lateral stratigraphy associated with the migration of numerous meander bends.

Braided channels or sinuous channels subject to frequent, large avulsions are likely to have much less predictable stratigraphy. Floodplains along these rivers can be described as a three-dimensional (3D) mosaic in which different grain sizes representing diverse depositional environments can be abruptly juxtaposed laterally and vertically (Figure S6.7). Arid-region braided channels, for example, repeatedly avulse across a broad floodplain. This led Graf (2001) to suggest that historical information about channel location—including floodplain stratigraphy—be used to develop a locational probability map that provides a statistical view of the river's most likely location at any point in time.

Lateral accretion of bars dominates floodplain stratigraphy of braided rivers, creating an irregular surface capped by a thin, discontinuous cover of finer, vertical accretion sediment accumulated in former channels (Dunne and Aalto, 2013). Levees along braided rivers tend to be less tall as a result of lateral channel movement. These levees are also easily eroded, so that sand splays are an important mechanism of floodplain sedimentation. Relatively steep gradients and sparse vegetation facilitate chute cutoffs and avulsions that can be rapidly filled if suspended sediment concentrations are high (Dunne and Aalto, 2013).

Although avulsion is particularly characteristic of braided rivers, this process can occur in any type of river. *Avulsion* refers to the shifting of a channel or channel belt across the floodplain. Avulsion typically occurs during floods, although downstream blockage of a channel by wood or ice, for example, can create sufficiently high water levels to promote avulsion in the absence of flooding. The new channel belt follows the zone of maximum floodplain slope while migrating to the lowest part of the floodplain. This process can occur over a period of years to centuries (Bridge, 2003), or during a single very large flood.

The frequency of avulsion increases with increasing deposition rate. Deposition enhances topographic relief and cross-valley slope relative to down-valley slope of the channel belt. Frequency of avulsion also increases with base level rise, which causes aggradation and growth of alluvial ridges (Bridge, 2003). Numerical simulations, and the age distribution of floodplain vegetation, suggest that mid-sized sinuous alluvial channels with sand and gravel beds preferentially reoccupy floodplain locations that were recently abandoned, resulting in a decreasing probability of floodplain reoccupation with time since abandonment (Konrad, 2012).

6.1.2 Erosional processes and floodplain turnover times

Floodplain sediment can be removed by localized erosion during extreme floods. Examples of localized erosion include the longitudinal grooves, scour marks, stripped floodplains, chutes, and anastomosing erosion channels described by Miller and Parkinson (1993) after a 1985 flood in West Virginia, USA. Longitudinal grooves are elongate linear grooves parallel or subparallel to the local direction of flood flow that extend down the valley floor tens to hundreds of meters, with depths and widths from a few centimeters to greater than 1 m. Miller and Parkinson (1993) interpreted these features as the early stages of floodplain erosion, because the rest of the valley bottom was largely unaltered. Scour marks vary in shape from small circular or elliptical pits to elongate parabolic or spindle-shaped and irregular marks that appear to be associated with flow separation and formation of vortices. Individual marks can be less than 0.3 m deep and 1 m in diameter, but the largest scour marks reach tens to hundreds of meters long and tens of meters wide. Stripped floodplains occur where flood waters remove a veneer of finer overbank deposits to reveal underlying cobble and boulder sediments across most or all of the floodplain. Concentrated flow on the floodplain produces a well-defined chute channel comparable in dimensions to the pre-flood river channel, rather than stripping the entire floodplain surface. Anastomosing erosion channels across the floodplain are associated with incomplete channel widening that leaves remnant islands in the expanded channel (Miller and Parkinson, 1993). Removal of floodplain vegetation by the flood strongly influences the ability of flood waters to create such localized erosional features.

Gradual change in floodplain morphology can facilitate abrupt, localized erosion during floods, a phenomenon Nanson (1986) described as catastrophic stripping. Studying partly confined floodplains along high-energy coastal rivers in New South Wales, Australia, Nanson described gradual overbank deposition over periods of 10^2–10^3 years that builds a floodplain of fine-textured alluvium with large levees. Growth of the levees and adjacent floodplain surfaces progressively restricts peak flows to the main channel and floodplain back-channels, concentrating erosional energy until widespread scour of the channel boundaries and floodplain occurs. This catastrophic stripping is discontinuous along a channel (Nanson, 1986).

Longitudinal variation in valley geometry and channel gradient can also result in localized floodplain erosion at the reach scale. Comparing the effects of widespread flooding in the northeastern United States, Patton (1988) found that sedimentation during large floods is important in constructing floodplains within highland drainages. In these drainages, reaches of lateral confinement and steep gradient experienced erosion during flooding. Less confined, lower-gradient reaches within the highland drainages experienced primarily deposition. Lowland drainages, in contrast, were not as strongly influenced by the flood, because magnitudes of flood erosion and deposition did not exceed magnitudes during a succession of more moderate flows.

More continuous floodplain erosion occurs as a result of channel lateral migration and bank erosion. Along the Brazilian portion of the Amazon River, sediment exchanges between the channel and floodplain in each direction exceed the annual flux of sediment out of the river at Óbidos (~1200 Mt/year) in the lower portion of the drainage (Dunne et al., 1998). The annual sediment supply entering the channel from bank erosion was estimated at 1.3× the Óbidos flux (1570 Mt/year). The annual volume of sediment transferred to the floodplain via channelized flow (460 Mt/year) and diffuse overbank flow (1230 Mt/year) totaled 1.7× the Óbidos flux. These estimates indicate a net accumulation of 200 Mt/year on the floodplain (Dunne et al., 1998). Quantitative studies of floodplain dynamics along the Strickland River of Papua New Guinea similarly indicate that ~50% of the total sediment load is recycled between the channel and floodplain via cut bank erosion and point bar deposition (Aalto et al., 2008). In gravel-bed channels such as the Fraser River of Canada, sediment inputs from bank erosion can exceed net downstream sediment flux by an order of magnitude, partly because bank sediment is largely transferred to adjacent bars, creating short sediment step lengths (Ham, 2005).

The continual exchange between sediment within the channel and sediment stored in the floodplain results in *floodplain turnover*, or erosion of existing floodplain sediment and deposition of new sediment (Figure S6.8). Turnover times typically increase downstream for two reasons. First, increasing floodplain width downstream means that longer time spans are needed for lateral channel migration to completely cross the floodplain. Second, even large floods are less likely to completely erode the floodplain than in smaller, laterally confined portions of the floodplain in upstream reaches (Patton, 1988). Mertes et al. (1996) estimated turnover times of ~1000 years in upstream portions of the Brazilian Amazon, and >4000 years in the downstream portions of the river near Óbidos. Based on cosmogenic ^{26}Al and ^{10}Be, Wittmann et al. (2011) inferred that some of the sediment in transport within the Amazon channel network has been stored in distal, deeply buried portions of floodplains for as long as 5 million years.

Because of the slow turnover times of sediment in many large floodplains, floodplain stratigraphy can provide a variety of paleoenvironmental information. The dimensions and stratigraphy of paleochannels can be used to infer past flow regimes (see Section 3.2.1). Pollen and fossils such as gastropods in the sediments filling abandoned floodplain lakes and channels can be used to infer past climates. Archeological sites are also common in floodplains and can provide useful information on floodplain chronology and on local plants and animals if food remains are present.

One of the implications of floodplain turnover is the potential for storage and subsequent release of contaminants. Several studies document preferential floodplain storage of mining-related contaminants. Floodplain storage results from dispersal of tailings during large floods caused by tailings dam failures (Marcus et al., 2001). Floodplain storage also reflects the tendency of mining-related contaminants such as mercury to travel adsorbed to the fine sediment that accumulates in floodplains via vertical accretion (Miller et al., 1999). Once deposited on the floodplain, the contaminants can slowly leak into the river via bank erosion over a period of centuries (Macklin et al., 1997).

6.1.3 Downstream trends in floodplain form and process

The general pattern of floodplain form is a progressive downstream increase in the longitudinal continuity and the lateral extent of floodplains. Exceptions can occur in low-relief terrains where even headwater channels have little lateral confinement, allowing wider, longitudinally continuous floodplains to develop. Exceptions also occur in rivers that cross mountain ranges. The Danube River of Europe, for example, heads in mountainous regions, but the middle and lower sections of the river's course alternate between wide alluvial basins with multi-thread channels and extensive floodplains, and narrowly confined bedrock gorges with little floodplain development (Wohl, 2011a).

Valley and floodplain width within regions of high relief can also exhibit substantial longitudinal variation as a result of local changes in bedrock erodibility, glacial history, sediment supply, network structure, and biota (Wohl, 2010). Changes in lithology and jointing can influence bedrock erodibility, so that a river alternates between wider, lower gradient segments where bedrock is more readily eroded and a floodplain develops, and steep, narrow gorges largely lacking floodplains where bedrock is more resistant (Ehlen and Wohl, 2002). Portions of a valley immediately upstream from a glacial moraine can be wider and of lower gradient, allowing more extensive floodplain development. Local point sources of coarse sediment, such as debris flows or landslides, can laterally constrict floodplains, as well as creating at least a temporary local base level that alters channel form (Korup, 2013). Network structure, as expressed in the arrangement and relative size of tributaries entering a main channel, influences water, sediment, and other fluxes (e.g., instream wood) into the mainstem, and affects the form and processes of the mainstem floodplain (Benda et al., 2004). Biota, including beavers that build dams and forests that allow sufficient wood recruitment to support channel-spanning logjams, can create channel segments of lower gradient and greater overbank flows (Westbrook et al., 2011; Polvi and Wohl, 2012).

Exceptions to the general downstream trend of increasing floodplain width can also result from structural controls or changes in substrate resistance. The Amazon River is entrenched where it crosses each of four structural arches, resulting in restricted channel movement and narrower floodplains through these segments (Mertes et al., 1996). Similar effects have been described along the Mississippi (Schumm et al., 1994) and other large alluvial rivers (Bridge, 2003). The very low gradients of large alluvial rivers allow these rivers to be modified by even slow, small tectonic movements (Dunne and Aalto, 2013), resulting in downstream changes in gradient, sediment transport, channel planform, and floodplain width and processes.

Regional- to continental-scale tectonic setting can also influence floodplain extent on very large rivers. The largest rivers with long, low-gradient reaches that facilitate extensive sedimentation and floodplain development drain the tectonically passive margins of continents (Potter, 1978). Rivers that head in orogens (belts of deformed rocks, typically associated with higher topographic relief) have greater sediment supplies than those that drain cratons (tectonically stable continental interiors, typically with lower topographic relief) (Milliman and Meade, 1983). If these rivers flow away from the orogen, like the Amazon, they are able to create extensive floodplains, whereas rivers that flow parallel to the orogen and receive sediment from tributaries along much of their length, as does the Orinoco, tend to have narrower and asymmetrical floodplains where tributary fans of coarse sediment shift the main river away from the orogen (Dunne and Aalto, 2013).

6.1.4 Classification of floodplains

Nanson and Croke (1992) suggested a three-part classification for floodplains, based on specific stream power and sediment texture as an influence on substrate resistance. The intent of this classification is to distinguish floodplains with respect to processes of origin and resulting morphology.

1. High-energy, non-cohesive floodplains with specific stream power >300 W/m^2 at bankfull are disequilibrium landforms that partly or completely erode during infrequent extreme floods. These floodplains typify steep upland portions of a drainage, in which bedrock or very coarse alluvium limits lateral channel migration. Relatively coarse vertical accretion deposits dominate these floodplains.
2. Medium-energy non-cohesive floodplains with bankfull specific stream power of 10–300 W/m^2 are in dynamic equilibrium with annual to decadal flow regime. Geomorphic change is limited during extreme floods because energy is dissipated by overbank flow. Lateral point bar accretion and braid-channel accretion tend to dominate deposition on these floodplains.
3. Low-energy cohesive floodplains with bankfull specific stream power <10 W/m^2 form along laterally stable channels with low gradients. Energy is dissipated by overbank flow during extreme floods and cohesive bank sediments limit lateral channel migration, so vertical accretion and infrequent channel avulsion dominate floodplain deposition.

Because floodplains do not necessarily progressively increase in width downstream, and because channel and floodplain characteristics are so closely related, some river classifications emphasize the degree of valley confinement and floodplain development. The River Styles framework of Brierley and Fryirs (2005), for example, begins by differentiating valley setting as being confined, partly confined or unconfined. Similarly, designations of river geomorphic spatial differentiation in mountainous settings emphasize the degree of valley confinement. Polvi et al. (2011) found that valley confinement and elevation explained nearly 90% of the variability in field-delineated width of the riparian zone, as defined based on plant species present, in headwater rivers of the Colorado Rocky Mountains.

A floodplain surface can remain active for thousands of years along a river with stable base level and relatively consistent water and sediment inputs. Along many rivers, however, the longevity of a particular floodplain surface is limited because incision or aggradation reduces the lateral connectivity between the channel and the floodplain surface. When this occurs, a terrace can form.

6.2 Terraces

Terraces are relict channel–floodplain features that have been isolated from contemporary river processes. Each terrace consists of a *tread*, the flat portion that represents the former floodplain surface, and a *riser*, the steep portion that separates the terrace from adjacent surfaces.

Terraces can be created by aggradation of the floodplain to a level no longer accessed by relatively frequent flows, or aggradation followed by incision. Terraces can also be created by incision of the channel. These various combinations can result from altered sediment supply or water flow caused by changes in climate, land cover, or tectonics. Terraces can also result from intrinsic processes associated with sediment transport that is discontinuous in time and space (Section 1.4). A change in base level or water or sediment discharge can lead to aggradation or degradation and terrace formation.

Terraces are not necessarily present along a river, but where they do exist, these landforms can be used to infer past longitudinal profiles of rivers, as well as paleoenvironmental conditions. Where terraces are present in a drainage basin, they typically do not extend fully into the headwaters because of lack of formation or subsequent erosion.

Terraces have been widely used by people because they provide flat, in many cases stable surfaces, which are close to rivers and suitable for agriculture and cities. Terraces containing coarse-grained channel-lag deposits are also mined for placer metals and construction aggregate.

6.2.1 Terrace classifications

Numerous classifications are used in connection with terraces. Some of these are simple and descriptive. *Paired terraces* have equivalent surfaces on both sides of a river valley. *Unpaired terraces* do not match across the valley and occur in situations such as continued incision while a sinuous river migrates laterally across the valley.

Other terrace classifications focus on the dominant process creating the terrace. The names of *erosional terraces* and *depositional terraces*, for example, are self-explanatory, although in practice it can be difficult to assign a primary cause of terrace formation because both erosion and deposition are involved in the formation of many terraces. Similarly, *tectonic terraces* and *climatic terraces* can be difficult to distinguish in a region with active tectonic uplift or relative base level fall that has also undergone climatically driven changes in water and sediment yield over the period of terrace formation.

One terrace classification distinguishes the duration of terrace formation. This classification differentiates *event terraces* associated with a single extreme perturbation such as a very large flood or debris flow, and *sustained terraces* that represent a persistent change in water and sediment yield.

One of the more straightforward classifications focuses on terrace composition and differentiates strath and fill terraces (Figure 6.3). *Strath terraces* have low-relief terrace treads formed in bedrock or other cohesive materials such as glacial till, overlain by a veneer of alluvium thin enough to be mobilized throughout its entire depth by the river. Existence of a strath terrace implies a period of vertical stability during which a river forms a relatively planar bedrock valley bottom by lateral erosion, followed by a period of vertical incision as transport capacity increases beyond sediment supply. Strath terraces are less likely to form as uplift and rate of incision increase, because the river is cutting downward too fast to form a strath tread (Merritts et al., 1994). Strath terraces tend to be more extensive where rivers flow over bedrock less resistant to weathering and erosion as a result of lithology or jointing, and more poorly developed over more resistant rock (Montgomery, 2004; Wohl, 2008a).

Fill terraces are alluvial sequences too thick to be mobilized throughout their depth by the river. Because fill terraces can form more rapidly than straths, they do not necessarily imply a period of vertical stability. They do imply a period of aggradation, followed by incision. Alluvium in fill terraces is commonly topped by 0.1–1 m of fine overbank sediments and eolian deposits (Pazzaglia, 2013).

A relatively common scenario among fill terraces associated with high-latitude climatic fluctuations during the Quaternary is illustrated by the example of the Maas River of northern Europe. Terrace alluvium aggraded along the Maas during periods of cold climate and glacial advance, whereas the river incised during periods of warm climate and glacial retreat (van den Berg and van Hoof, 2001). Alternating glacial and interglacial periods during the Quaternary produced 21 paired and unpaired fill terraces and ~100 m of incision along the Maas (van den Berg and van Hoof, 2001).

6.2.2 Mechanisms of terrace formation and preservation

Terraces can be interpreted as deviations from a graded condition (Section 5.4). A graded river can aggrade in response to base level rise or incise in

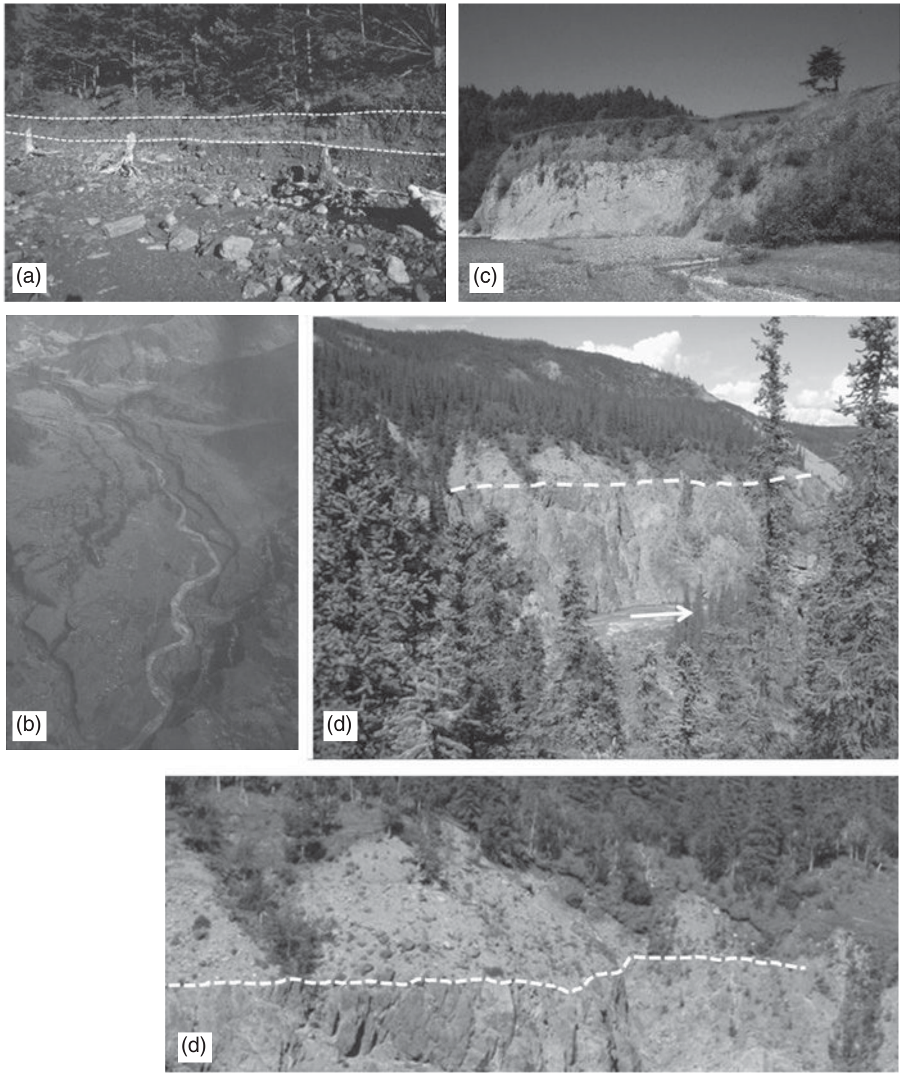


Figure 6.3 Examples of river terraces. (a) Two small fill terraces (top outer edge of each terrace indicated by dashed white line) are visible in this view of a small creek in northern California, USA. Successive waves of aggradation and incision along this creek created the higher fill terrace just visible at the base of the forest, then the younger fill terrace about 1 m in elevation below the upper terrace, and finally resulted in lowering and widening of the channel, killing the riparian trees that left the stumps visible here. Flow in the channel is from left to right. (b) Aerial view of numerous fill terraces along a river in Nepal. (c) A strath terrace along the Mattole River in California, USA. The white material exposed in the lower portion of the cut bank is bedrock. An upward fining alluvial sequence of cobbles to loamy soil overlies the bedrock strath. Flow is from the foreground toward the rear in this view. The cut bank is approximately 6 m tall. (d) Distant view of very tall strath terrace along the Duke River in northwestern Canada. Strath surface is approximately 100 m above the active channel. Lower view shows close-up of bedrock-fill contact, which is indicated by dashed line in both photographs. Arrow indicates flow direction on main channel.

response to base level fall, as long as the channel remains fixed to the changing base level such that the river maintains a steady-state profile with uniform valley and channel geometry and constant concavity and steepness (Gilbert, 1877; Mackin, 1937; Pazzaglia, 2013). Creation and preservation of a terrace reflects unsteadiness in the profile because the channel aggrades and/or incises, and geometry, concavity, and steepness can be altered. This unsteadiness is a transient response to changes in water and sediment input or base level.

Strath terraces have been attributed to a wide variety of changes in river dynamics (Wohl, 2010; Pazzaglia, 2013), including

- periods of balanced sediment supply;
- altered sediment supply;
- glacial–interglacial transitions;
- tectonically induced changes in rock uplift;
- falling local base level, eustatic base level fall (global sea level change); and
- autocyclic oscillations in erosion rate in laterally migrating channels.

Numerical modeling suggests that formation of strath terraces depends on input variability that creates a changing ratio of vertical to lateral erosion rates (Hancock and Anderson, 2002). A wide variety of external and internal factors can create input variability, which explains how strath terraces can result from different causes in diverse river catchments or within a catchment through time.

Similarly, fill terraces have been attributed to diverse changes (Wohl, 2010; Pazzaglia, 2013), including

- fluctuating water and sediment discharge during glacial cycles, volcanic eruptions, climatic change, and changing land use;
- repetitive lateral shifting and stillstands during continuous downcutting;
- fluctuating base level;
- complex response to base level change; and
- sediment waves migrating down catchments over periods of tens to hundreds of years in response to hillslope mass movements or land use.

Fill terraces differ from strath terraces in that fill terraces can be event-based features resulting from a single flood, debris flow, or landslide that overwhelms sediment transport capacity and causes floodplain aggradation before river flow subsequently incises the deposit (Miller and Benda, 2000; Strecker et al., 2003). Fill terraces are also more likely to be reduced in planform size during lateral erosion after the terrace forms (Moody and Meade, 2008).

The mechanisms by which terraces form, along with terrace geochronology, have been the subject of much research because a terrace can record some change in external control variables, such as water and sediment yield or base level, and thus provide paleoenvironmental information. Terrace studies began in the nineteenth century with the conceptual model that river terraces reflected an externally imposed change in river dynamics associated with glaciation, uplift, or changing base level (Davis, 1902b). Terraces continued to be studied under this assumption until Schumm's work on semiarid alluvial channels (Schumm and Hadley, 1957) and physical experiments (Schumm, 1973) demonstrated that terraces could result from crossing internal thresholds while external inputs to the river network remained unchanging. Schumm and Hadley (1957) described a cycle of erosion for ephemeral rivers in semiarid regions that involves longitudinally discontinuous incision. Infiltration into the streambed during brief periods of flow causes downstream decreases in discharge, local aggradation, steepened channel gradients, and incision, all without any change in climate or land cover that alters water and sediment inputs to the river network. This represented a major shift in thinking about how some types of terrace could form, and led to the idea that event-based terraces could result from temporary fluctuations in water and sediment supply without major shifts in climate, tectonics, or land use.

Supplemental Section 6.2.2 reviews physical experiments and numerical simulations used to understand mechanisms of terrace formation and preservation.

6.2.3 Terraces as paleoprofiles and paleoenvironmental indicators

Grain size and stratigraphy of terraces can be used to infer sediment supply and flow conditions (Section 3.2.1). The longitudinal profile of a terrace can approximate the paleoprofile of the river. Age of the terrace can be used to constrain the timing of events that resulted in terrace formation. However, inferring paleoenvironmental conditions from terrace characteristics is complicated by several factors.

First, a given change in climate that alters water and sediment yield to a river network can create very different responses within the river network, depending on the climatic conditions present prior to the change. This is illustrated by Bull's (1991) work in the Charwell River basin of central New Zealand. Different portions of the river basin responded differently when regional precipitation increased slightly and temperature increased more substantially. The lower portion of the drainage, from the basin mouth to the contemporary tree line, changed from periglacial conditions to greater vegetation density and soil thickness. This reduced sediment yield, increased peak discharge, and resulted in stream incision. The upper portion of the drainage above timber line experienced an increase in periglacial processes and increased sediment yield, so that channels in this zone did not incise (Bull, 1991).

Second, a change in base level, such as that caused by tectonic uplift, can result in different responses across a river network. For example, larger rivers in the vicinity of a triple junction in northern California, USA, can maintain uniform incision into bedrock during relatively rapid uplift, creating sequences of strath terraces along the rivers (Merritts and Vincent, 1989; Merritts et al., 1994). Smaller, headwater rivers cannot incise rapidly enough to keep pace with uplift, and instead become steeper with time rather than creating strath terraces. Studies elsewhere indicate that, even on larger rivers, terrace formation can be longitudinally discontinuous if a river lacks the power to create strath surfaces along segments of more resistant bedrock (Wohl, 2008a).

Third, event-based terraces that result from local disturbances such as landslides create fill terraces of limited longitudinal extent (Montgomery and Abbe, 2006; Wohl et al., 2009). Even more sustained changes such as base level rise or fall may only cause terraces in the lower portion of a river network (Merritts et al., 1994). Strath terraces, in particular, may require some minimum drainage area to form (Garcia, 2006) because of the stream energy required to create a strath surface in bedrock. The formation and preservation of straths can also vary with differences in bedrock erosional resistance (Montgomery, 2004; Wohl, 2008a). Terraces may be thickest, widest, and best preserved near tributary junctions where tributary sediment widens the mainstem and helps to build a terrace sufficiently large to be preserved (Pazzaglia, 2013).

Fourth, a single perturbation such as base level fall can result in multiple terraces if complex response occurs (Section 6.1.1; Schumm, 1973). Each terrace along a river therefore cannot necessarily be correlated with an external change.

Fifth, the duration of time represented in terrace alluvium can vary widely. The alluvium capping a strath terrace is synchronous with formation of a strath, for example, whereas the alluvium of a fill terrace is younger than the underlying surface, and can accumulate over long periods of time (Pazzaglia, 2013).

Sixth, the timing of terrace formation can substantially lag the timing of any external perturbation in water and sediment yield or base level (Pazzaglia, 2013). Base level on the Drac River in the French Alps dropped 800 m following removal of an ice-dam during glacial retreat. As a knickpoint moved upstream through the basin in response to base level lowering, terrace formation lagged behind base level drop by 2000–5000 years, depending on location within the drainage (Brocard et al., 2003).

These complicating factors do not by any means render terraces useless as paleoenvironmental indicators. The presence of spatial and temporal differences and lag times in terrace formation within a river network only means that terraces must be interpreted carefully and in a broader context than a single terrace exposure, as with any other river feature. An example of paleoenvironmental interpretation comes from Merritts et al. (1994), who interpreted Quaternary strath and fill terraces along three rivers in northern California in the context of river response to tectonic uplift and fluctuating base level associated with changes in sea level. Fluctuating sea level strongly influenced terraces along the lower portion of each river (tens of kilometers long). Aggradation led to formation of fill terraces during sea level high stands. Incision occurred during low stands. Long-term uplift created unpaired strath terraces in the middle and the upper portion of each river.

Strath and fill terraces can be used to reconstruct river longitudinal profile at the time of strath or floodplain formation (Paola and Mohrig, 1996), although creation of straths, in particular, can lag by several thousand years the input changes that cause their formation (Merritts and Vincent, 1989). Both types of terraces also can subsequently be tectonically deformed (Finnegan, 2013). The terrace surface is the easiest proxy to use for paleoprofile. However, this surface can be deformed by weathering and inputs of colluvium from adjacent hillslopes, so the more difficult-to-discern underlying bedrock strath provides a better indicator of original longitudinal profile (Pazzaglia, 2013).

Any paleoenvironmental inferences drawn from terraces rely on establishing a chronology of terrace formation (Wohl, 2010; Pazzaglia, 2013). Terrace chronology can be based on absolute methods. Absolute methods provide a numerically precise age, although the accuracy of this age is not necessarily greater than that achieved using relative methods. Absolute methods applied to terrace ages include

- ^{14}C dating of organic matter in terrace sediments (Merritts et al., 1994);
- dendrochronology of terrace vegetation (Pierson, 2007);
- luminescence techniques (Pederson et al., 2006) (luminescence measures energy stored in sediment as a result of natural, background radioactive decay—the energy stored is a function of time that the sediment has been buried, as well as background decay rate);
- U-series dating of pedogenic carbonate in terrace sediments (Candy et al., 2004);
- cosmogenic isotope dating of terrace sediments or bedrock in strath terraces (Riihimaki et al., 2006);
- paleomagnetic dating of terrace sediments (Pan et al., 2003);
- tephrochronology of volcanic ashes incorporated in terrace sediments (Dethier, 2001); and
- radiometric ages of bounding lithologic units such as basalt flows (Maddy et al., 2005).

Relative methods infer differences in age between terraces by characterizing parameters that change through time, although the rate of change in a parameter is not necessarily linear. Relative methods applied to terrace chronology include

- development of weathering rinds on coarse clasts in terrace sediments (Pazzaglia and Brandon, 2001);
- soil characteristics (thickness, accumulation of translocated clays or soluble salts, color, and iron-oxide speciation) (Tsai et al., 2007);
- amino acid racemization of organic compounds such as shells (Penkman et al., 2007); and
- lichenometry (the diameter of lichen colonies, which increases with time) (Baumgart-Kotarba et al., 2003).

Many studies utilize multiple techniques (Guralnik et al., 2010). The actual event or period of time dated can be either the phase of tread construction (erosion of a strath or deposition of a fill) or the incision that produces a terrace scarp (Ritter et al., 2011). The phase of tread construction is time transgressive and of longer duration, whereas incision can be widespread and rapid (Pazzaglia, 2013).

As numerous detailed case studies accumulate in the scientific literature on fluvial geomorphology, diverse groups have begun to establish digital catalogs and databases. The Fluvial Archives Group (FLAG) is actively documenting the global record of terrace stratigraphy (Vandenberghe and Maddy, 2000). FLAG creates databases for Pleistocene and Holocene sedimentary archives.

As with other types of fluvial deposits, LiDAR and DEMs facilitate designation and mapping of terraces (Jones et al., 2007). Shallow geophysical techniques such as ground penetrating radar and electrical resistivity (Froese et al., 2005) are used to image terrace subsurface geometry. These images can substantially enhance information available from limited surface exposures.

6.3 Alluvial Fans

Alluvial fans are primarily depositional environments that are shaped by the combined effects of erosion and deposition, similar to the other features addressed in this chapter. Channels exiting high gradient, laterally confined canyons in regions with high relief commonly create a fan-shaped deposit known generically as an *alluvial fan*, although colluvial processes can contribute to sediment accumulation. Water flow, hyperconcentrated flow, debris flow, rockfall, landslide, and snow avalanches can all contribute sediment to alluvial fans. Fans can be differentiated based on primary depositional process and morphology, resulting in debris cones dominated by rockfalls, debris fans dominated by debris flow, and so forth. Fans classified in this way can be differentiated as those dominated by *inertial transport* (debris flows, rockfalls) versus those dominated by *traction transport* (hyperconcentrated and fluvial flows) (Stock, 2013).

Where parallel channels drain from a mountain range into a basin, adjacent alluvial fans can coalesce to form an *alluvial apron* or *bajada*. *Fluvial megafans* are unusually large alluvial fans that mostly occur between 15° and 35° north and south of the equator where rivers that undergo moderate to extreme discharge fluctuations enter actively aggrading basins (Leier et al., 2005). Australians use the term *floodout* to describe sites where a marked reduction in channel capacity creates greater overbank flows and associated deposition (Tooth, 1999). Floodouts bear some similarities to fans, but are not necessarily associated with a change in lateral channel confinement. *Paraglacial fans* are alluvial fans composed mainly of reworked glacial deposits (Ryder, 1971). *Fan deltas* form at the margins of the ocean, where rivers flow from a mountainous region across a narrow coastal zone, and can grade into deltas that form in freshwater or marine subaqueous environments (Section 6.4). Small alluvial fans form at crevasse splays below levees on floodplains.

Alluvial fans can form in any climatic region. However, fans in arid and semiarid environments are by far the most well studied, not least because the lack of continuous vegetation cover results in good exposure of fan features (Figures 6.4, S6.9, and S6.10).

Lower transport capacity than upstream channel reaches promotes aggradation, overbank flows, and frequent avulsion on alluvial fans, which creates numerous hazards for structures built on these surfaces. Fans are also heterogeneous with respect to process, form, grain size and stratigraphy, and soils and vegetation. Fan sediments tend to be relatively porous and permeable, which facilitates infiltration and subsurface flow, allowing fans to hydrologically buffer a catchment (Vivoni et al., 2006). Herron and Wilson (2001), for example, described how an alluvial fan in the subhumid temperate climate of southeastern Australia significantly buffered hydrology in a 26 ha catchment by storing 20%–100% of surface runoff delivered to the fan. Wetlands can occur at the downstream end of fans in humid climates as lower gradients promote deposition of finer, less permeable sediments and subsurface flow rises back toward the surface (Woods et al., 2006).

Fans can also buffer sediment delivery by at least temporarily storing sediment. Fans thus lower the connectivity of sediment movement and the coupling between hillslopes and channels or between tributaries and larger rivers. The volume of sediment stored can fluctuate substantially through time, as processes of fan erosion—incision of the main or distributary channels, or erosion of the toe of a tributary fan by a larger river—alternate with processes of deposition. In the Grand Canyon of Arizona, USA, for example, flash floods and debris flows along small, steep, ephemeral channels create tributary fans that laterally constrict the mainstem

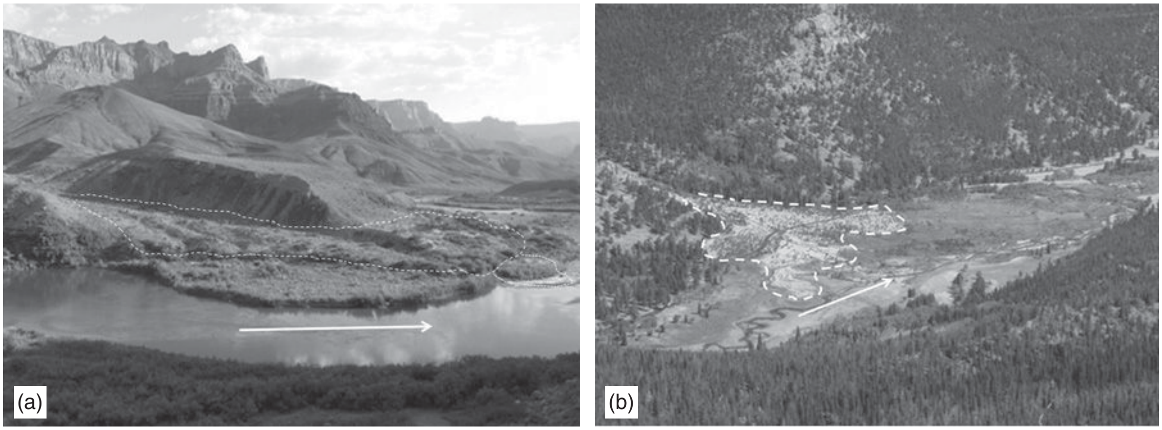


Figure 6.4 Examples of alluvial fans. (a) A fan created at the junction of a tributary with the Colorado River in the Grand Canyon, USA. The older fan deposits are outlined with the dashed white line, the most recent fan deposits with the dotted white line at far right center. Vegetated deposits in the foreground are the floodplain. Mainstem flow is left to right. (b) An event-based alluvial fan created by a damburst flood down the Roaring River in Colorado, USA. The Roaring River enters the Fall River valley: the Fall River flows from lower left to upper right in this view. The outburst flood occurred in 1982, and the resulting alluvial fan appears lighter in color (and outlined in dashed white line) in this photograph, taken nearly 30 years later, because of slow growth of vegetation on the fan surface. See also Figure S6.4 for additional, color photographs of fans.

Colorado River. Fans continue to grow until the associated lateral constrictions create supercritical flow during large floods on the mainstem, when the fans are eroded sufficiently to allow mainstem flow to become subcritical (Kieffer, 1989).

6.3.1 Erosional and depositional processes

Alluvial fans are geomorphically complex and dynamic environments. Channelized water flows, hyperconcentrated flows, and debris flows alternate spatially and temporally with unconfined sheet flow or with rapid infiltration that results in sieve deposits and subsurface flow (Griffiths et al., 2006). Deposits on many fans are coarse grained and poorly sorted as a result of relatively short transport distances, inputs by debris flows and flash floods, and rapid loss of flow capacity as a result of infiltration or overtopping of shallow channels (Blair and McPherson, 2009).

Episodic aggradation at the fanhead causes oversteepening that initiates incision. Consequently, an incised channel, or *fanhead trench*, is present near the apex of many alluvial fans, with active deposition concentrated farther down the fan (Blair and McPherson, 2009). Channels also frequently avulse or are abandoned through stream capture.

Downstream from the zone of most active incision, channels are commonly braided and can decrease in size downstream as a result of discharge infiltration, so that channelized flow can give way to sheetflow (Bridge, 2003). Although deposition can produce relatively smooth surfaces, secondary channel networks can dissect these surfaces once deposition shifts elsewhere on the fan.

Blair and McPherson (2009) distinguished primary and secondary depositional processes. Primary processes transport sediment from the catchment onto an alluvial fan and cause fan construction. Secondary processes modify previously deposited sediment via overland flow, wind erosion or deposition, bioturbation, soil development, weathering, and toe erosion. Secondary processes dominate fan surfaces away from the usually limited areas of recent deposition.

Blair and McPherson (2009) also distinguished 3 classes and 13 different types of alluvial fans

(Table S6.1) based on dominant primary processes and textures. The three classes (bedrock, cohesive colluvium, non-cohesive colluvium) reflect the catchment slope material from which primary processes are triggered.

As Stock (2013) noted, very few direct observations of transport and depositional processes on fluvially dominated fans have been published. These observations consistently indicate

- fluctuations in water and sediment supply during a single storm,
- formation of coarse bars by particle accretion and subsequent diversion of flow and sediment around these bedforms on gravel-bed surfaces—these processes of channel bifurcation and avulsion are analogous to those in braided rivers,
- shallow, hydraulically rough flow, and unstable flows,
- critical flow with standing waves and presumably antidunes in sand- to cobble-bed surfaces, and
- large concentrations of suspended and bedload transport.

Processes of sediment accumulation seem to be consistently associated with bifurcation, avulsion, and surges of water and sediment on alluvial fans.

Alluvial fans are typically built over time by continuing deposition, but individual fans can result from a single extreme event, such as massive erosion associated with a damburst flood. The Roaring River alluvial fan in Colorado, USA, (Figure 6.4b) was created by a dam failure on a headwater lake in 1982. The resulting outburst flood flowed more than 7 km down the steep, narrow Roaring River, eroding substantial quantities of sediment from the Pleistocene-age glacial tills that underlie the river course. Where the Roaring River enters the wider, lower gradient Fall River valley, the flood created an alluvial fan covering 0.25 km^2 and up to 14 m thick (Blair, 2001), which persists more than 30 years later.

Fans can also be dominated by a specific and sometimes repetitive type of disturbance, such as tributary glaciation, volcanic eruption, wildfire-induced mass movements (Figure S6.11), and landslides (Korup et al., 2004; Torres et al., 2004). An example comes from the Southern Alps of New Zealand, where Korup et al. (2004) compiled case studies documenting a series of alluvial fans dominated by sediment inputs from large (>10^6 m^3) landslides.

Supplemental Section 6.3 describes approaches used to measure and model processes on alluvial fans.

Complex and intermittent erosion and deposition on alluvial fans create special challenges for hazard zoning and mitigation. Fans are commonly the only gently sloping surfaces available in high-relief environments, and hazards associated with erosion and deposition are intermittent in time. Consequently, urbanization concentrates on fans (Zorn et al., 2006). Approaches used to mitigate hazards (Kellerhals and Church, 1990; Pelletier et al., 2005; Wohl, 2010; Stock, 2013) include

- channelization and bank stabilization to limit overbank flows, sheetwash, and channel avulsion;
- debris interception barriers and detention basins to contain sediment entering the fan, rather than allowing the sediment to disperse and bury infrastructure or fill active channels;
- hazard mapping to guide development of infrastructure; and
- warning devices to facilitate rapid evacuation in case of an erosional or depositional event.

6.3.2 Fan geometry and stratigraphy

An alluvial fan approximates a segment of a cone radiating downslope from a point where a channel flows out from a high-relief catchment (Blair and McPherson, 2009). In plan view, a fan is arcuate, creating a 180 degree semicircle unless the toe of the fan is truncated by erosional processes, or the sides of the fan are restricted by adjacent fans. Longitudinal fan profiles within and away from the channel are typically segmented, with the steepest slopes along the flanks (Hooke and Rohrer, 1979). Segmentation can reflect intermittent uplift, channel incision, and episodic deposition of various types

(Bull, 1964; Arboleya et al., 2008). Channel gradient typically decreases downstream from 0.10–0.04 m/m to 0.01 m/m on fluvially dominated fans (Stock et al., 2007). This decrease in gradient reflects progressive deposition and declining sediment loads and grain size down-fan. Radial fan profiles tend to be concave or planar, and cross-fan profiles convex (Bull, 1977; Blair and McPherson, 2009). Fans typically extend 0.5–10 km from a mountain front, but large fans can extend 20 km (Blair and McPherson, 2009).

The volume of sediment available for deposition on a fan and the processes of deposition exert first-order controls on fan geometry. Climate, lithology, tectonics, and drainage area all influence sediment delivery to a fan. Climate influences weathering and caliber of sediment supplied to the fan, as well as flow available to transport sediment. Oguchi and Ohmori (1994) compared alluvial fans formed in the wet climate of Japan against fans formed in the arid American Southwest, across a range of lithologies. They found that, although fan area in each region is proportional to the catchment sediment output per unit time, sediment yield per unit area is higher in the Japanese basins for the same basin slope.

Lithologies that weather to very coarse sediment deposited by mass movements typically create the steepest fans. Lithologies that weather rapidly and release very fine sediments result in moderately steep fans of larger volumes, whereas lithologies that weather to sand-sized sediment create low-gradient fans of smaller volume per unit drainage area (Bull, 1964; Blair, 1999).

Tectonics influence rates of sediment production and delivery to rivers in upstream catchments, and the accommodation space in depositional sites. A basin that is subsiding relative to the source area can develop a thicker fan of more limited spatial extent, whereas a tectonically stable basin can develop a thinner but more areally extensive fan (Whipple and Trayler, 1996).

Tectonics can dominate fan formation in regions with active extension (divergence of plates), compression (convergence of plates), or transtension (lateral plate movement) in which confined, steep valleys are juxtaposed with unconfined depositional regions such as piedmonts (Stock, 2013). These areas favor large fans because of the lack of confinement on depositional space. Extensional regions such as the US Basin and Range, Tibetan Plateau, and Andean back-arc are characterized by mountain ranges and intervening subsiding basins along the margins of which alluvial fans form. These regions tend to be arid because of rain-shadow effects from bounding mountain ranges and current global circulation patterns, which facilitates identification and study of their alluvial fans. Some of the largest fans, such as the Kosi fan, form in convergent tectonic settings such as along the Himalayan or Taiwan thrust fronts (Stock, 2013). In this setting, active uplift provides abundant sources of sediment, but the depositional area can be less constrained than in an extensional setting with numerous mountain ranges. Other examples of formerly tectonically active settings conductive to fan formation are retreating escarpments such as South Africa's Drakensberg or the southeastern Australia passive margin, which juxtapose sediment sources in higher-relief areas with relatively flat piedmonts.

In some of the earliest quantitative work on alluvial fans, Bull (1962) found that drainage area corresponds to fan size (A_f) and fan slope (S_f) as

$$A_f = aA_d^n \tag{6.1}$$

$$S_f = cA_d^d \tag{6.2}$$

with a varying slope of 0.7–1.1 for the regression line (n), which averages 0.9 for fans in the United States. Coefficient a varies by more than an order of magnitude as a result of lithology, climate, and the space available for deposition on the fan (Viseras et al., 2003). Although these equations continue to be widely applied, Blair and McPherson (2009) noted limitations. These include the comparison of plan-view areas for 3D features (Equation 6.1), which are not of constant thickness and for which thickness is commonly unknown, and the lack of clear guidelines for defining area, which may not be straightforward (Stock, 2013).

Various morphometric indices, including relief and ruggedness, as well as particle size and roundness, and fabric of deposits, can be used to distinguish spatial differences in depositional processes on fans (de Scally and Owens, 2004; Stock, 2013) and to distinguish dominant depositional processes on individual fans (Al-Farraj and Harvey, 2005). Debris flows with high sediment concentration can create steep and rugged topography on the upper fan, for example, whereas low sediment concentration can create smoother topography on the lower fan surface by filling channels and other depressions (Whipple and Dunne, 1992).

Spatial and temporal heterogeneity in depositional processes on alluvial fans make the resulting stratigraphic records challenging to interpret. Nonetheless, an extensive literature, summarized in Wohl (2010), documents how changes in the volume and style of fan deposition through time have been used to infer, for example,

- climatically driven changes in sediment supply to alluvial fans in Japan (Oguchi and Oguchi, 2004; Waters et al., 2010), and changes in sediment supply associated with alpine glacier retreat in Switzerland (Hornung et al., 2010);
- human-induced and tectonically driven changes in sediment supply in the Solway Firth–Morecambe Bay region of Great Britain (Chiverrell et al., 2007) or deposition in the Tagliamento River region of Italy (Spaliviero, 2003);
- the magnitude and frequency of event-based sedimentation associated with storms or wildfires in the US Northern Rocky Mountains (Pierce and Meyer, 2008); and
- to construct short- and long-term sediment budgets for the upstream catchment of the Storutla River in southern Norway (McEwen et al., 2002).

An example comes from the work by Pierce and Meyer (2008) in Yellowstone National Park in Wyoming, USA. Much of the park is covered in conifer forests that burn at regular intervals, mostly as a result of lightning strikes during periods of drought. Pierce and Meyer used tree-ring records of fire and drought, pollen and charcoal in lake sediments, and charcoal and other sediments in numerous alluvial fans across the park to infer changes in fire regime and sedimentation over the past 2000 years. Alluvial fan sediments provided a longer record (8000 year) of fire than tree rings, were ubiquitous in this mountainous environment, and recorded stand-replacing fires that limited tree-ring records. Although alluvial fan deposition is episodic in time and space, discontinuities can be offset by compiling records from numerous individual stratigraphic sections (Pierce and Meyer, 2008).

6.4 Deltas

A delta is analogous to an alluvial fan in that it is primarily a depositional feature that results from a loss of transport capacity: in the case of a delta, where a river enters a body of standing water. Just as an alluvial fan is roughly triangular in shape, the word delta was first applied to a fluvial deposit 2500 years ago by the historian Herodotus because of the Nile delta's resemblance to the Greek letter delta, Δ (Ritter et al., 2011).

A delta can protrude well beyond the adjacent coastline when a river carries large volumes of sediment, and currents in the receiving body have limited ability to rework that sediment. Examples of deltas that protrude well beyond the adjacent shore include the Mississippi River delta in the Gulf of Mexico, the Danube in the Black Sea, the Nile in the Mediterranean Sea, Italy's Po–Adige delta in the Adriatic Sea, Russia's Volga River delta in the Caspian Sea, the Ganges–Brahmaputra and Irrawaddy in the Bay of Bengal, and Africa's Niger in the Gulf of Guinea.

Or the delta deposits can be primarily upstream from the adjacent coastline, so that the delta does not protrude into the ocean as a result of limited particulate sediment transport from the river basin or thorough reworking and transport of fluvial sediment by currents in the receiving water body. Examples include the Amazon and the Rio de la Plata in South America, Australia's Murray–Darling, the Congo, and the Columbia in North America.

Regardless of the shape, deltas are biologically diverse and productive environments that typically support commercial fisheries. Deltas also host vegetative communities that can help to reduce the velocity and associated damage of incoming oceanic storm surges. Delta features such as area of islands and bars and type and extent of vegetation influence the *tidal prism*—the volume of water exchanged in a tidal cycle—as well as values of bed shear stress and thus sediment dynamics (Canestrelli et al., 2010).

As river flow entering the body of standing water expands and decelerates, sediment is deposited. Various types of currents in the receiving water body rework the river sediment, sometimes with the result that no delta forms. The morphology and stratigraphy of the delta reflect the balance between river inputs and reworking of sediment by the receiving water body. In general, the coarsest sediment is deposited closest to the river mouth and finer sediment in suspension is carried farther into the receiving body. Large databases containing data from numerous deltas indicate that the area of a delta is best predicted from average discharge, total sediment load entering the delta, and offshore accommodation space (Syvitski and Saito, 2007).

Many of the world's great cities are located on deltas along marine shorelines. This is typically a mixed blessing for cities such as Shanghai, New Orleans, Calcutta, Rotterdam, Marseille, and Lisbon. On the plus side, deltas provide ease of transport and navigation of goods from the terrestrial interior and across the ocean, fertile soils and rich fisheries, and reserves of oil and natural gas in delta sediments. These benefits are offset by some of the same hazards common on alluvial fans, such as overbank flooding and channel avulsion. Deltas also include hazards associated with

- subsidence of delta sediments—an example comes from the Pearl River delta of China, where measurements based on interferometric synthetic aperture radar data indicate average subsidence of 2.5 mm/year within 500 m of the coast, and maximum values of 6 mm/year associated with rapid urban development, in a region less than 2 m above current mean sea level (Wang et al., 2012);
- flooding of low-lying portions of the delta during marine storm surges;
- marine erosion of the delta if river sediment supply decreases or sea level rises—deltaic coastlines in Greece, for example, are predicted to retreat 700–2000 m inland under projected sea level rises of 0.5–1 m as a result of combined effects of inundation during sea level rise and associated coastal erosion (Poulos et al., 2009);
- loss of water supply, transport, or waste disposal if river flow shifts location across the delta; and
- salinization of delta soils and groundwater if river flow decreases, sea level rises, or groundwater is pumped at a rate faster than natural recharge—in southeastern Spain, for example, increasing water demand for crop irrigation and reduced irrigation return flow linked to new irrigation techniques have reduced recharge of some coastal aquifers, which have become more saline since 1975 (Rodríguez-Rodríguez et al., 2011).

Like alluvial fans, deltas exemplify *terra* ***non-*** *firma*.

6.4.1 Processes of erosion and deposition

Deposition on deltas, as on alluvial fans, constantly shifts location across or down the delta in response to changes in fluvial water and sediment discharge, as well as changes in delta morphology resulting from deposition, subsidence, and reworking by waves and tides. The classic model of delta deposition involves deposition of a longitudinal bar where river transport capacity declines because of mixing between river water and the water of the receiving basin. The initial bar forces the river to bifurcate, and the process continues, developing a *distributary channel* network. Distributary channels formed at least in part by avulsion are lined with natural levees. With continued deposition, the distributary network progrades into the receiving body.

Slingerland and Smith (1998) developed a model predicting avulsion stability based on the ratio of the water surface slopes of the two branches of a bifurcation. Ratios < 1 (where the secondary channel slope is the numerator and the main channel slope is the denominator) result in a failed avulsion in which the incipient avulsion fills with sediment.

Ratios > 5 create avulsions in which the river abandons its original course in favor of the steeper channel.

Jerolmack and Mohrig (2007) proposed that avulsion frequency scales with the time needed for sedimentation on the streambed to create a deposit equal to one channel depth. They used relative rates of bank erosion and channel sedimentation to derive a dimensionless mobility number to predict the conditions under which distributary channels form.

An active delta includes *subdeltas* created when sediment diverted through breached levees accumulates as crevasse splays. Channels on the subdelta also bifurcate and prograde until they develop gradients similar to those of the main channel and deposition shifts to a new subdelta. On the Mississippi River, the process of forming and abandoning subdeltas occurs over a few decades (Morgan, 1970). Physical experiments suggest that the timescale for these processes on river-dominated deltas reflects the time over which a river mouth bar that is prograding basinward reaches a critical size and stops prograding (Edmonds et al., 2009). This triggers a wave of bed aggradation that moves upstream, increasing overbank flows and shear stresses exerted on the levees, triggering avulsion and the growth of subdeltas.

In high-latitude rivers, ice and ice-jam floods can exert an important control on processes of delta erosion and deposition. Canada's Mackenzie River provides a well-studied example. The Mackenzie, which drains 1.8 million km^2, enters the Arctic Ocean at 70° N. The Mackenzie Delta stretches across 12,000 km^2, including a 200 km maze of meandering channels and 45,000 riparian lakes (Goulding et al. 2009b). High spring flows from snowmelt dominate Mackenzie River runoff. The northerly flow direction of the river results in the spring flow peak progressing downstream with the seasonal advance of warm weather. The flood wave commonly encounters an intact and resistant ice cover downstream, causing the formation of ice jams and enhanced overbank flooding (Prowse and Beltaos, 2002). The spatial extent and duration of overbank flooding influence erosion and deposition; habitat abundance, diversity, and connectivity; biogeochemical processes; lake flooding; and even surface albedo as floodwaters hasten the melting of snow and ice, causing a slight increase in air temperatures at a critical time for the growth of terrestrial and aquatic plants (Goulding et al., 2009a). Ice transported by the floodwaters into overbank areas enhances localized erosion. Decreases in the severity of river-ice break-up during recent decades have lessened the flooding of some delta lakes, which may lead to loss of these water bodies, as well as changes in the biogeochemical interactions between river water and the floodplain ecosystem, and the processes described earlier (Prowse et al., 2011).

As a delta front progrades further into the receiving basin, shorter routes for water flow into the basin become available along the sides of the delta. These shorter routes can also access topographically lower portions of the coastline than the delta, which has been progressively built up as well as out. River flows commonly access such shorter routes via crevasses developed in levees well upstream from the delta, causing *delta switching*, a relatively abrupt shift in river course and delta location. Over time, delta switching results in distinct depositional lobes within the general depositional complex (Aslan and Autin, 1999). Delta switching occurs approximately every 1000–2000 years along the Mississippi River (Aslan and Autin, 1999), for example, and along the Ebro River in Spain (Somoza et al., 1998).

6.4.2 Delta morphology and stratigraphy

Deltas reflect the changing balance through time between at least four factors (Bridge, 2003).

1. The volume and grain-size distribution of river sediments: The characteristics of the river sediments supplied to the delta integrate everything in the river catchment that influences sediment supply—lithology, relief, tectonics, climate, land cover, and land use—as well as river discharge and transport capacity. The greater the sediment load coming down the river, the more likely the river is to build a delta. Conversely, the stronger the

currents in the receiving water body, the more likely these are to rework the river sediments and dominate the size and form of the delta, or to completely remove the river sediments.

2. The relative density of the river water and the receiving water body: The relative densities of river and receiving waters reflect salinity, temperature, and concentration of suspended sediment. When the densities are similar (*homopycnal flow*), the river and basin water mix in three dimensions closer to the shoreline and much of the sediment is deposited close to the river mouth (Bridge, 2003). This is most likely to occur where a river enters a freshwater lake. River water denser than basin water (*hyperpycnal*)—such as very cold or turbid water entering a freshwater lake—will flow as a bottom current, mixing only along its lateral margins, carrying sediment farther into the basin. Hyperpycnal flows can also occur where a river enters an ocean, as on the Huang He (Yellow River) of China. River water of lower density than the basin water (*hypopycnal*)—such as freshwater entering saline water—can flow over the denser water as a buoyant plume, mixing along the bottom and sides, and carrying sediment further into the basin than in the case of homopycnal flow. This type of flow is common on the delta of the Mississippi River.
3. The currents of the receiving water body: Lakes and enclosed seas typically experience lower-energy waves and tidal currents, although storm-generated winds can temporarily enhance wave energy. Deltas in these lower-energy environments are likely to reflect predominantly fluvial processes. River mouths entering long, narrow embayments such as the Bay of Fundy on Canada's southeastern coastline, or semi-enclosed seas, such as the North Sea in the Atlantic or the Sea of Cortez off northwestern Mexico, can have very strong tidal currents that cause tidal bores to travel up the bay and into the entering river, strongly influencing delta shape and size.
4. Subsidence and water-level change in the receiving body, which influence relative change in base level: Changes in relative base level influence river gradient and transport capacity, the location of the shoreline, and the energy of currents in the receiving basin. Either rise or fall of relative base level can shift the locations of active erosion and deposition across a delta. An example comes from a Pleistocene-age lake delta in northwestern Germany, in which periods of lake-level rise correspond to vertically stacked delta systems as depositional centers shifted upslope, and lake-level fall corresponds to development of a single incised valley with coarse-grained delta lobes in front of the valley (Winsemann et al., 2011).

One of the more commonly used classifications of deltas emphasizes the relative importance of fluvial processes or reworking by the receiving body (Figure 6.5) (Ritter et al., 2011). *High-constructive deltas* reflect predominantly fluvial deposition; these are sometimes known as *river-dominated deltas* (Fagherazzi, 2008). River-dominated deltas can be *elongate*, with greater length (in the downstream direction) than width, like the contemporary delta of the Mississippi River. Elongate deltas have more silt and clay and subside rapidly once deposition ceases or shifts elsewhere on the delta. River-dominated deltas can also be *lobate*, with greater width than length, and slightly coarser sediment that subsides more slowly upon abandonment, allowing time for waves and tides to rework the sandy sediment. The now-abandoned Holocene deltas of the Mississippi River are lobate.

High-destructive deltas are shaped by currents in the receiving water body. Sediment is moved alongshore to form beach ridges and arcuate sand barriers in *wave-dominated deltas* such as those of the Rhone or Nile River. *Tide-dominated deltas* have strongly developed tidal channels, bars, and tidal flats, with sand deposits that radiate linearly from the river mouth, such as those in the Gulf of Papua (Ritter et al., 2011).

Any type of delta can be subdivided into delta plains and a delta front. The majority of most deltas consists of *delta plains*, extensive areas of low slope crossed by distributary channels. Delta plains include subaerial portions of floodplains, levees, and crevasse splays, and subaqueous portions

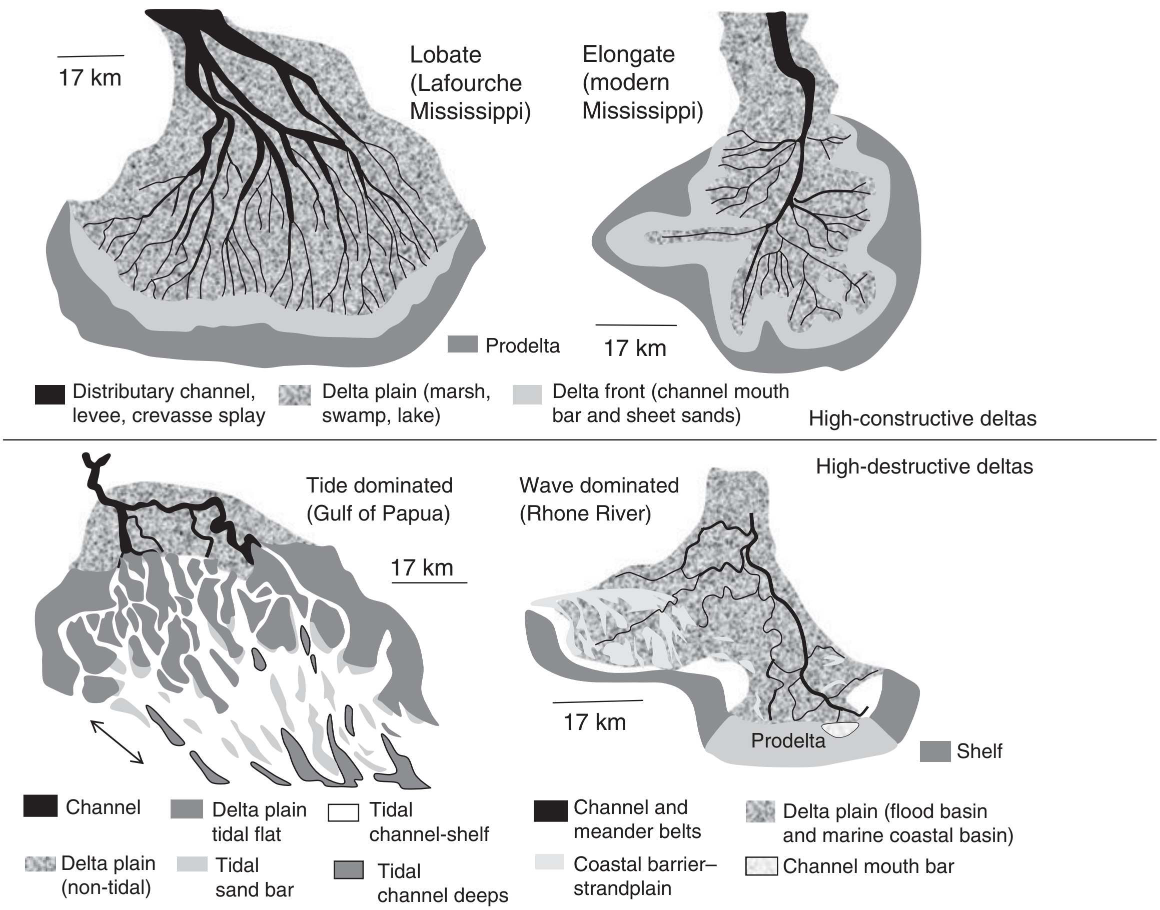


Figure 6.5 Basic delta types. (Adapted from Ritter et al., 2011, Figure 7.35.)

of marshes, lakes, interdistributary bays, tidal drainage channels, and tidal flats (Bridge, 2003). The gradient of the delta plain correlates with the ratio of sediment supply to sediment retention on the delta, sediment concentration (a proxy for delta plain sedimentation), and mean water discharge (Syvitski and Saito, 2007). The delta plain gradient increases as sediment accumulates on the plain, but the gradient decreases as river discharge increases.

Delta plains can have extensive vegetation cover that varies with climatic setting, rates of deposition, and—down the delta—salinity gradients. Some of the world's great deltas are covered in dense subtropical forest and mangrove swamps (e.g., the Sundarbans at the mouth of the Ganges River). Other deltas have primarily sedges, rushes, and marsh plants (e.g., Mississippi River delta, Pearl River delta in China), rather than woody vegetation. Regardless of the specific plant communities, vegetation cover strongly increases the resistance of delta sediments to erosion. Removal of vegetation typically results in much faster and more severe delta erosion, as well as less efficient trapping and storage of incoming sediment. Analysis of historical aerial photographs of the Skagit River delta in Washington, USA, for example, suggests that sand bars formed within tidal channels are colonized by marsh vegetation, which forms a marsh island that narrows the gap between the island and the mainland marsh. This constricts the tidal channel and eventually leads to coalescence of the island with the adjacent mainland (Hood, 2010).

Delta plains are more affected by river currents than by wave and tidal currents, and thus resemble inland floodplains (Bridge, 2003). Deposition along the lower portions of tributary channels can be enhanced during periods of high tides or storm surges, leading to increased frequency of avulsion (Bridge, 2003).

The *delta front* is the steeper, basinward portion of a delta, where sediment-carrying river water enters open water. High rates of deposition and limited reworking by currents in the receiving basin can result in steepening of the delta front beyond the angle of repose of the sediment, causing a sediment gravity flow that can develop into a turbidity current. A *turbidity current* is analogous to a subaqueous debris flow, in that it is a relatively dense, flowing slurry of sediment that follows bottom topography and can travel farther into the receiving basin than suspended sediment. Oceanographers first documented turbidity currents when slumps initiated by the 1929 Grand Banks earthquake were transformed into a turbidity current that flowed down the slope of the Atlantic Ocean basin, breaking submarine telegraph cables en route (Heezen et al., 1954).

Deposition on a delta front is episodic, as is deposition on the basin floor beyond the delta. Stratigraphy in a prograding delta that is progressively building into the receiving basin is characterized by *bottomsets*—fine-grained, low inclination deposits of the basin floor; *foresets*—deposits of the delta front at the angle of repose; and *topsets*—relatively flat-lying and coarse deposits of the delta plain (Bridge, 2003) (Figure 6.6). G.K. Gilbert (1885) introduced these terms, so deltas with this type of stratigraphy are known as *Gilbert-type deltas*. Edmonds et al. (2011) refer to Gilbert-type deltas as *foreset-dominated deltas*, and distinguish them from *topset-dominated deltas* in which the distributary channels incise into pre-delta sediment.

As with other fluvial processes, delta dynamics can be studied using physical experiments (Edmonds et al., 2009) and numerical simulations (Geleynse et al., 2011). Canestrelli et al. (2010) provided an example of a numerical simulation used to describe the propagation of the tidal wave within the delta and along the channel of the Fly River in Papua New Guinea. Using a high-resolution bathymetric map of the delta and a two-dimensional (2D) finite element model, they examined the sensitivity

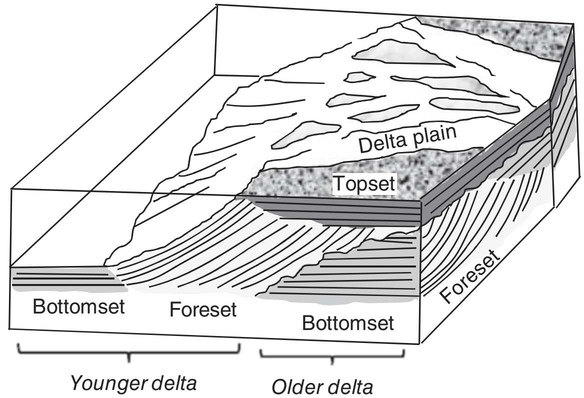


Figure 6.6 Schematic block diagram of a Gilbert-type or foreset-dominated delta, showing the three primary types of beds, the delta plain, bars, and distributary channels. River flow enters a body of standing water from the right.

of water fluxes and bed shear stresses to varying sea level and the area covered by delta islands. Model results suggest that a decrease in island area or an increase in sea level results in increased tidal prism and bed shear stresses.

Fagherazzi and Overeem (2007) provided a thorough review of numerical models of delta dynamics. Numerical models include:

- 2D cross-shelf models that focus on the evolution of the continental shelf profile and ignore along-shelf variations in morphology;
- 2D along-shelf models, which focus on the evolution of the along-shelf morphology and consider either a constant cross-shelf profile or integrate across the shelf;
- pseudo-3D shelf area models based on depth-averaged formations of hydrodynamics and sediment transport; and
- fully 3D models that reproduce the full structure of the flow field coupled with sediment transport.

6.4.3 Paleoenvironmental records

The depositional environment that is a delta integrates upstream changes in yields of water, sediment, and other materials. Consequently, delta morphology and stratigraphy are commonly examined as a source of information on catchment paleoenvironments. The length of paleoenvironmental record that can be accessed in this manner partly depends on the age and erosional history of the delta (in the sense of erosion removing portions of the depositional record), and partly on the techniques used to examine deltas. Changes in sedimentary facies, mineralogy, and isotopic ratios, as well as changes in location of delta sublobes, are used to infer changes in depositional processes driven by either upstream changes in water and sediment yield or changes in coastal wave and tide energy (Hori et al., 2004; Hori and Saito, 2007; Liu et al., 2010). Chronologies of delta deposition can be constrained using luminescence and radiocarbon techniques (Hori and Saito, 2007; Tamura et al., 2012).

The Nile River delta provides a nice example (Krom et al. 2002). One of the Nile's major tributaries, the White Nile, drains predominantly crystalline rocks. Other major tributaries—the Blue Nile and the Atbara—drain the Tertiary volcanic rocks of the Ethiopian Highlands. The different rock types have different ratios of $^{87}Sr/^{86}SR$ and Ti/Al. Changes in these ratios through time, combined with known changes in river discharge, indicate that periods of higher river flow during the past 7000 years correspond to decreased input of total sediment and sediment derived from the Blue Nile watershed because of higher rainfall and increased vegetative cover in the Ethiopian Highlands (Krom et al., 2002).

Over shorter time periods, changes in delta morphology and process can be studied using aerial photographs, satellite images, and repeat ground-based surveys. Tamura et al. (2010), for example, used topographic maps and satellite images to study changes in the shoreline of the Mekong River delta since 1936 (Tamura et al., 2010). They found that the delta has changed asymmetrically over this period in response to net southwest-ward transport of sediment associated with winter monsoons.

6.4.4 Deltas in the Anthropocene

As with other aspects of river networks, many deltas have recently undergone major changes as a result of direct and indirect human alterations of water and sediment yield to the delta, and of the delta distributary network. As noted earlier, many deltas are densely populated and heavily farmed—close to half a billion people live on or near deltas, and an estimated 25% of the world's population lives on deltaic lowlands (Tamura et al., 2012). These people are increasingly vulnerable to changes in delta morphology and dynamics.

One of the primary changes is sediment compaction and delta subsidence, so that deltas are sinking more rapidly than sea level is rising (Syvitski et al., 2009). Delta subsidence results from several factors. Sediment consolidates with time as pore water is expelled. Trapping of sediments and river flow in reservoirs upstream from the delta reduces ongoing inputs that would otherwise offset progressive compaction and subsidence. Removal of oil, gas, and water from delta sediments reduces pore pressure and facilitates compaction. Floodplain engineering on the delta, including channelization, limits overbank deposition and conveys sediment more efficiently beyond the delta and into the ocean. And, rising sea level subjects deltas to more intense erosion by wave and tidal currents, as well as storm surges.

In China's Changjiang (Yangtze) delta, for example, construction of approximately 50,000 dams throughout the contributing watershed has resulted in erosion of the delta (Yang et al., 2011). Numerous other examples of reduced sediment yield to deltas and associated erosion have been documented (Jun et al., 2010; Nageswara Rao et al., 2010). Diversion of water and sediment flows either away from the delta or to a different portion of the coastline exacerbates delta shifting, causing the abandoned delta to compact, subside, and erode (Jabaloy-Sanchez et al., 2010; Restrepo and Kettner, 2012).

Examining 33 deltas around the world, Syvitski et al. (2009) found that 85% of the deltas experienced severe flooding during the previous decade, causing the temporary submergence of 260,000 km^2. They also conservatively estimated that delta surface area vulnerable to flooding could increase by 50% under current projections of sea level rise during the twenty-first century.

Coastal wetlands associated with deltas are also being lost to submergence or erosion. A 2005 survey of 42 deltas across the globe indicated the loss of nearly 16,000 km^2 of wetlands during the preceding 20 years, with an average rate of loss of 95 km^2 per year (Coleman et al., 2005).

Mining of construction aggregate, land reclamation, and changes in delta morphology can also influence the pathways of water and sediment across the delta, as well as delta submergence and the ability of waves and tides to rework delta sediments. Large-scale sand excavation from China's Pearl River delta since the mid-1980s, for example, has resulted in increased water depth and lowered streambed elevation (Zhang et al., 2010). This has led to an increased tidal prism and upstream movement of the tidal limit, which have facilitated increased saltwater intrusion into the estuary.

Some deltas changed little during the twentieth century and their aggradation rate remains in balance with, or exceeds, subsidence or relative sea-level rise. Examples include deltas on the Amazon (Brazil), Congo (Democratic Republic of Congo), Orinoco (Venezuela), Fly (Papua New Guinea), and Mahaka (Borneo) Rivers (Syvitski et al., 2009). These deltas typically occur on rivers with lower density of human settlement in the upstream drainage basin and minimal flow and sediment regulation. Deltas with minimal human manipulation provide a control against which to evaluate the effects of direct human alteration of deltas within the context of changing climate and sea level.

Conversely, the Pearl Delta in China and the Mekong Delta in Vietnam are among the most at-risk deltas. Each of these deltas is inhabited by millions of people, is exposed to typhoons, and has limited coastal barrier protection. Each of these deltas also has compaction and/or reduced aggradation, so that much of the delta surface area is below mean sea level (Syvitski et al., 2009).

Human-induced changes are also exacerbated where very little or no river flow now reaches the delta as a result of consumptive water uses upstream. Deltas subject to this severe change in flow regime include those on the Indus River and the Colorado River in the United States and Mexico.

In addition to changes in morphology caused by human alterations, delta sediments can be locally contaminated by toxic synthetic chemicals such as pesticides (Feo et al., 2010) or petroleum byproducts (Wang et al., 2011). Any portion of a river network can be affected by such contaminants, but deltas are particularly vulnerable because they integrate diverse upstream sources of contamination. Contaminants can affect the delta environment and, when remobilized by wave or tidal erosion, can be spread into the nearshore environment. Naturally occurring contaminants, such as arsenic associated with highly reducing deltaic sediments in the Bengal and Mekong River deltas, can also strongly influence delta environments (Berg et al., 2007).

6.5 Estuaries

An *estuary* represents an incised river valley along the coastline of an ocean (Figure S6.12). The river valley incised during a period of lower sea level that caused base level lowering for the river, typically during a period of continental ice sheet formation. Subsequent sea level rise during ice sheet melting and retreat backflooded the estuary. The typical scenario is that the estuary grows progressively shallower and less subject to marine influence as terrestrial and marine sediments accumulate. An estuary that is a sediment sink is also referred to as a *microtidal estuary* (Cooper, 2001). Usually, estuarine filling occurs during and/or shortly after backflooding on sediment-rich rivers. Estuary filling can require time periods longer than the Holocene on rivers with very low sediment loads, in settings in which river floods periodically scour the estuary, or in settings in which sediment is primarily transported through the estuary to the marine environment (Cooper, 2001; Cooper et al., 2012).

Numerous factors influence the form and processes of estuaries (Dalrymple and Choi, 2007). Among the most fundamental are:

the bathymetry of the estuary, which ranges from a relatively shallow-water, channelized environment landward of the coast, to deeper, unconfined settings on the shelf, the source of the energy that moves sediment, from predominantly river currents to tidal, wave and oceanic shelf currents,

the frequency, rate and direction of sediment movement, which vary from

- unidirectional, continuous movement to seasonal or episodic when driven by rivers,
- reversing in tidal settings,
- episodic and coast-parallel when driven by waves, and
- onshore–offshore in tide-dominated shelf environments, and

water salinity, which ranges from fresh, through brackish, to fully marine.

Given this breadth of river and coastal influences, estuaries exhibit significant diversity. Most estuaries, however, possess a three-part structure composed of (i) an outer, marine-dominated portion where the net bedload transport is headward, (ii) a relatively low-energy central zone with net bedload convergence, and (iii) an inner, river-dominated (but still marine-influenced) portion where net sediment transport is seaward (Dalrymple et al., 1992). Each zone is developed to differing degrees among diverse estuaries as a reflection of site-specific sediment availability, gradient of the coastal zone, and the stage of estuary development.

Various classifications exist for estuaries. Examining estuaries in South Africa, for example, Cooper (2001) distinguished five types of estuaries based on contemporary morphodynamics. *Normally open estuaries* maintain a semi-permanent connection with the open sea. The three primary categories of open estuaries are barrier-inlet systems maintained by river discharge (river-dominated estuaries) and tidal discharge (tide-dominated estuaries), and open estuaries that lack a supratidal barrier because of inadequate availability of marine sediment. *Closed estuaries* are separated from the sea for long periods by a continuous supratidal barrier. Perched closed estuaries develop behind high berms and maintain a water level above high tide level, whereas non-perched closed estuaries develop behind barriers of lower elevation with wide beach profiles.

Dalrymple et al. (1992) distinguished simply wave- and tide-dominated estuaries. *Wave-dominated estuaries* typically have three well-defined portions: a marine sand body composed of barrier, washover, tidal inlet and tidal delta deposits; a fine-grained central basin; and a bay-head delta that experiences tidal and/or salt-water influences (Figure 6.7). Examples of wave-dominated estuaries include the Hawkesbury Estuary and Lake Macquarie in Australia, eastern shore estuaries in Nova Scotia and the Miramichi River in Canada, and San Antonio Bay and Lavaca Bay in the United States (Dalrymple et al., 1992). *Tide-dominated estuaries* have a marine sand body composed of elongate sand bars and broad sand flats; a central zone of tight meanders where bedload is transported by flood-tidal and river currents; and an inner, river-dominated zone with a single, low-sinuosity channel. Examples of tide-dominated estuaries include the Adelaide and Ord Rivers in Australia, the Cumberland Basin and Avon River in Canada, and Alaska's Cook Inlet in the United States (Dalrymple et al., 1992).

Estuaries share some characteristics with deltas in that they include depositional surfaces on which tidally influenced channels form. Numerous studies demonstrate that these tidal channels are funnel or trumpet shaped, such that channel width tapers upstream in an approximately exponential fashion (Chappell and Woodroffe, 1994; Fagherazzi and Furbish, 2001; Savenije, 2005). The shape of the exponential width profile relates to the mouth width and river output. Channels and estuaries with larger mouths have a more pronounced funnel shape than systems with narrower mouths, and higher river water and sediment discharge create a longer "funnel" (Davies and Woodroffe, 2010).

As in the case of deltas, human coastal settlement and resource use concentrate around estuaries. Estuaries play an important role in the cycling of carbon and other nutrients through the exchange and modification of terrestrial organic matter transported by rivers to the ocean (Canuel et al., 2012). Because estuaries typically retain organic matter for some period of time, these environments have high rates of primary production and support abundant

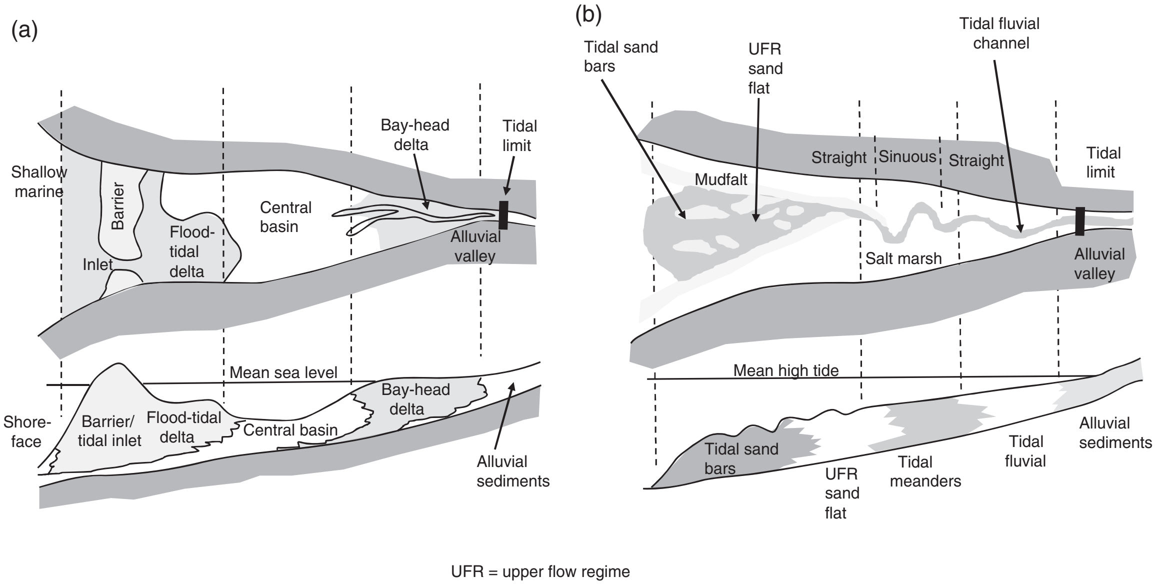


Figure 6.7 Distribution of morphological elements in plan view and sedimentary facies in longitudinal section within an idealized (a) wave-dominated and (b) tide-dominated estuary. For (a), the barrier/sand plug is shown as attached to the headland, but may not be connected on low-gradient coasts, where it can be separated from the mainland by a lagoon. The longitudinal section represents the onset of estuary filling after a period of sea level rise. For (b), the longitudinal section is along the axis of the channel and does not show all sedimentary facies present. (From Dalrymple et al., 1992, Figures 4 and 7.)

and diverse wetlands and fisheries. Dramatically increased fluxes of nitrogen to estuaries as a result of livestock wastes, industrial-scale application of agricultural fertilizers, and other human activities have caused many estuaries to experience eutrophication (Green et al., 2004). *Eutrophication* is an ecosystem response to increased nutrients. This response typically involves a substantial increase in phytoplankton and associated decreases in dissolved oxygen, which together cause *hypoxia*, or the so-called dead zones. At least 375 hypoxic coastal zones have been identified around the world, primarily around western Europe, the eastern and southern coasts of the United States, and Asia (CENR, 2000).

Many of the same techniques used to study deltas are applied to the study of process and form in estuaries. Bathymetric mapping of submerged features is particularly useful in studying estuaries, and navigational charts for individual estuaries can date from the early nineteenth century (van der Wal and Pye, 2003).

6.6 Summary

The fluvial landforms discussed in this chapter—floodplains, terraces, alluvial fans, deltas, estuaries—are inherently dynamic and rapidly changing environments. Ever-shifting balances among different processes of deposition and erosion can reflect changes external to the landform such as variation in climate or base level. Shifting erosion and deposition can also reflect internal adjustments within the environment of the landform while the external setting remains relatively stable, as when continued development of levee crevasses results in delta shifting. This continual variability can make fluvial landforms difficult to interpret, but floodplains, terraces, fans, and deltas, in particular, are nonetheless valuable repositories of information on river dynamics over time spans of tens to tens of thousands of years.

Human activities increasingly alter process and form in these extra-channel environments. Indirect effects associated with climate change and rising sea level combine with direct effects including flow regulation, levee construction, and channelization to change fluxes of water, sediment, solutes, and organisms between channels and extra-channel environments. Altered fluxes—altered connectivity—have in many cases led to reduced water quality, lower biological productivity and ecological sustainability, and increased hazards to human populations and infrastructure as a result of eroding landforms. The great challenge of the future is to understand how past human actions have changed river networks, and how undesirable changes can be mitigated or reversed as we go forward.

Chapter 7
Humans and rivers

Each of the preceding sections of this book has included some mention of how people alter drainage basin and channel process and form. Because human influence is now so ubiquitous and, in many cases, intensive, this chapter focuses specifically on several aspects of human activities that exert an important influence on river process and form. One of the definitions for "impact" in the Oxford English dictionary is the "strong effect of one thing, person or action, on another; an influence." This is the sense in which this chapter discusses human impacts on rivers.

Human impacts can be indirect or direct. *Indirect human impacts* on rivers result from activities that occur outside of the river network and do not directly alter channel form or process. Human-induced changes in atmospheric chemistry, and the resulting changes in atmospheric and oceanic circulation, and temperature and precipitation patterns, are one category of indirect impacts that have received increasing attention within the past two decades. These changes have been underway for more than a century, but are receiving widespread attention only now. Land cover is strictly defined as the type and density of vegetation, although alterations in vegetation are commonly accompanied by alterations in topography and soil characteristics. Changes in land cover have been occurring for thousands of years, and recognition of the effects on river networks goes back more than a century (Marsh, 1864; James and Marcus, 2006).

Direct human impacts result from activities within the river network that directly alter channel form or process. Flow regulation, and specifically the construction of dams, has received the most scientific and public scrutiny, in part because individual dams can be extremely large and can alter physical, chemical, and biological characteristics of entire watersheds. Other, typically smaller-scale alterations of channel form and connectivity—associated with flow diversion, construction of levees, dredging and channelization, bank stabilization, and instream mining of placer metals and sediment for construction aggregate—can have cumulative effects equally substantial to those associated with large dams. Most studies of these impacts, however, focus on individual or regional examples, and there is not yet an effective analysis of the cumulative, global effects of these activities.

The third portion of the chapter examines river engineering designed specifically to reverse or mitigate the negative consequences of earlier river engineering. River restoration or rehabilitation attempts to alter river process and form in a manner perceived as beneficial for fisheries, recreation, esthetics, or some other characteristic not considered in prior engineering. Instream flows, channel maintenance flows, and environmental flows are a form of

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river engineering that focuses primarily on the flow regime as a means of protecting or restoring river process and form.

The chapter concludes with a discussion of river health, as defined from ecological and geomorphic perspectives. River health is an intuitively appealing concept that evaluates the cumulative effects of numerous human-induced alterations in river ecosystems and provides a conceptual framework for undertaking river restoration.

Humans impact landscapes through the effects of economic activities, including management intended to mitigate natural hazards, but landscapes also impact humans through patterns of resource availability and natural hazards. Werner and McNamara (2007) modeled these interactions as hierarchical complex systems that provide insight into how human manipulation of landscape processes both influence and respond to landscape process and form.

The concept of landscape sensitivity (Brunsden and Thornes, 1979) is also important in thinking about humans and rivers. A *sensitive landscape* is likely to respond to a change in external controls. A *resilient landscape* is likely to return to initial conditions following a disturbance. Whether examining how rivers respond to climate change, deforestation, or river restoration, the idea that some river networks or some portions of a river network are more sensitive or resilient (Section 4.3.2) is critical to understanding and predicting river adjustment (James and Lecce, 2013). River segments with very high sediment transport capacity may adjust relatively quickly and easily to increased sediment yield, for example, whereas portions of the network that are transport limited may exhibit substantial change in channel geometry in response to a similar magnitude of increased sediment yield.

7.1 Indirect impacts

7.1.1 Climate change

Systematic sampling of atmospheric gases during the past half century, as well as air bubbles trapped in polar ice sheets, record increasing atmospheric concentrations of CO_2 and other compounds that permit a given volume of air to absorb more infrared radiation. Increased CO_2 levels have caused measurable increases in air temperature since the start of systematic measurements, and particularly since the 1950s. International scientific panels estimate that continued increases in CO_2 will cause a rise in average global temperature during the twenty-first century of anywhere from 1° to 5°C (IPCC, 2008). Increases in atmospheric temperature cascade through atmospheric, oceanic, freshwater, and terrestrial systems as changes in numerous characteristics of these systems (Table 7.1; Figure 7.1). Even a brief consideration of the extent and complexity of local to regional changes in river process and form associated with global changes in air temperature reveals the difficulty of predicting and adapting to these changes on more than a relatively crude, basic level.

For river networks, changes in precipitation associated with changing atmospheric temperature are the most immediate manifestation of climate change. Most studies of the effects of climate change on rivers consequently start with an examination of how altered precipitation characteristics will likely influence a river's flow regime. Predictions of future precipitation commonly rely on general circulation models (GCMs). At present, these models typically have a spatial resolution based on grid cells as small as 12,100 km^2. They can thus be used for examining large river basins, but smaller spatial scales require either statistical down-scaling to relate local climate variables to large-scale meteorological predictions (Gyalistras et al., 1997; Andreasson et al., 2003), or a more detailed regional model that is nested within the GCM (Giorgi et al., 1994; Marinucci et al., 1995). Inferred precipitation characteristics must be coupled with simulations of soil moisture, land cover, infiltration, and runoff, as discussed in Chapter 2.

Examples of the complexity of such modeling efforts come from studies in mountainous regions, where even a small increase in air temperature can significantly influence the distribution, volume, snow water equivalent, and snowmelt timing of mountain snowpacks (López-Moreno et al., 2009;

Table 7.1 Examples of how changing global climate will impact rivers.

Description	Reference
Because temperature and precipitation thresholds influence the magnitude and frequency of mass movements, climate change is increasing sediment yields to mountain rivers in the Himalayan and Tibetan Plateau	Lu et al. (2010)
In the semiarid/arid Western United States, changes in air temperature, moisture, and thunderstorms alter the frequency and severity of wildfires, which causes changes in water, sediment and nutrient yield to streams, as well as changes in forest structure and wood recruitment to streams; predictions include loss of floodplain forests	Hauer et al., 1997 Westerling et al. (2006) Rood et al. (2008)
In Alaska, USA, warming increases glacial runoff, which changes flow regime, water temperature and sediment discharge, causing changes in channel substrate, bedforms, channel stability, leaf litter quantity and quality, and habitat complexity	Oswood et al. (1992)
Along arid-region rivers, changes in flood magnitude and frequency can cause channel incision that removes the deep hyporheic sediments that support microbial communities: changes in precipitation and runoff can also alter the availability of nitrogen, which is a limiting nutrient in these rivers	Grimm and Fisher (1992)
As the proportion of precipitation coming in the form of rain (rather than snow) increases at high elevations, winter rains increase flood hazards and decrease groundwater and summer streamflow by up to 50% in the mountains of central Europe	Eckhardt and Ulbrich (2003)
Droughts in the United Kingdom during the late twentieth century highlighted the sensitivity of water resources to climatic fluctuations; predictions suggest that rainfall is likely to become more intense and less frequent in future	Wilby (1995)
In Australia, changes in rainfall will create the greatest runoff changes in arid catchments; some regions will receive more runoff (more intense and frequent rainfall in eastern Australia), others will receive less; frequency and severity of droughts will also increase	Chiew and McMahon (2002)
Changes in temperature and hydrology that promote disturbances to vegetation such as wildfire, insect outbreaks, and drought-related die off are predicted to increase sediment yield to rivers in the northern US Rocky Mountains; increased sediment yields will affect aquatic habitat and downstream water-storage reservoirs	Goode et al. (2012)
Increasing air temperatures in the monsoonal Mahanadi River basin of India during the twentieth century correlate with declining river flows	Rao (1995)

Gillan et al., 2010). Nonlinear responses complicate attempts at modeling and prediction. Snowpack on a glacier, for example, reduces absorbed solar radiation and melt rate (Oerlemans and Klok, 2004). Removal of the snowpack can increase daily discharge amplitude by more than 1000% (Willis et al., 2002). Upward retreat of rainfall–snow elevation limits may thus substantially increase rates of glacial melting. Changes in glacial mass balance are particularly important in mountainous regions because even relatively small alpine glaciers provide large reserves of freshwater that sustain summer peak flows and autumn base flows. In populous, relatively dry lowlands such as parts of South America, India, and Pakistan, disappearance of mountain glaciers will substantially reduce water supply and likely cause social upheaval (Hasnain, 2002; Barnett et al., 2005; Singh et al., 2006).

Changes measured during the past few decades can provide insight into such multi-faceted alterations. Shifts in large-scale atmospheric circulation have resulted in altered flood frequency across Switzerland (Schmocker-Fackel and Naef, 2010). Earlier snowmelt, reduced snow accumulation, and altered streamflow are particularly well documented for the European Alps, Himalaya, and western North America (Kundzewicz et al., 2007). Regional-scale studies document how differences in characteristics

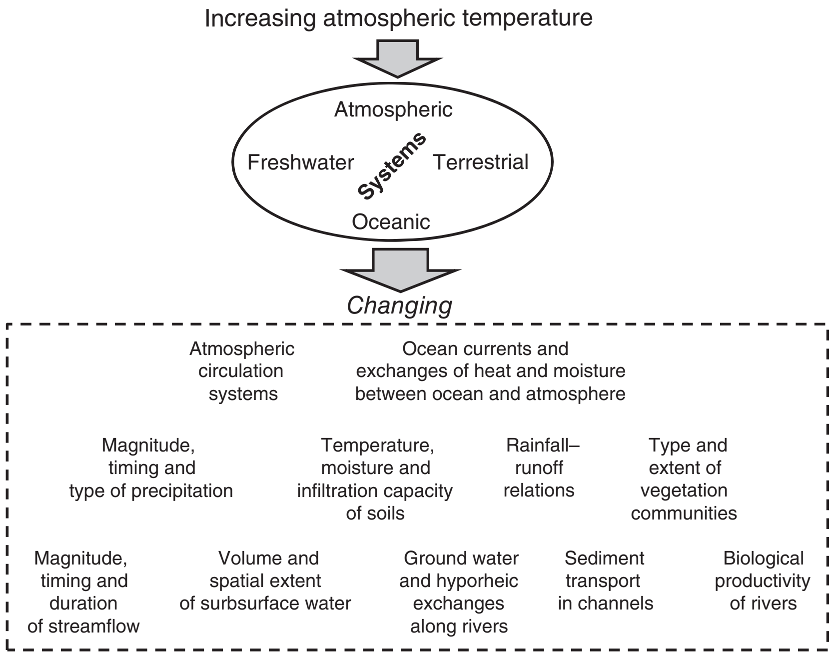

Figure 7.1 Schematic illustration of changes affecting rivers as a result of increasing atmospheric temperature.

such as subsurface drainage and seasonal distribution of precipitation influence between-watershed differences in sensitivity of streamflow to climate warming. In the Oregon Cascade Range of the western United States, for example, differences in bedrock geology between watersheds correspond to differences in volume and seasonal flux of subsurface water (Tague et al., 2008; Tague, 2009). Watersheds dominated by extensive, low-relief basaltic lava flows have deeper groundwater flow and are predicted to show greater absolute reduction in summer streamflow under expected temperature increases than are watersheds dominated by shallow subsurface flow, where stream flow is already flashy and highly seasonal.

At the broadest scale, modeling suggests that precipitation totals will increase in high latitudes and in mid-latitudes during winter. Precipitation variability and possibly the frequency of extreme precipitation will increase in the tropics. Snow cover duration will decrease and mid-latitude soil moisture may decrease in summer (Kattenberg et al., 1996; Mote et al., 2005). The largest hydrological changes are predicted for snow-dominated basins of mid-to-higher latitudes (Nijssen et al., 2001). Understanding and preparing for the implications of these changes at the scale of individual drainage basins requires a great deal more effort, but it is worth emphasizing that hydroclimatic changes that influence river ecosystems are not hypothetical. Many regions of the world already exhibit statistically significant differences between the latter half of the twentieth century and earlier periods (Marchenko et al., 2007; Van Der Schrier et al., 2007; Rood et al., 2008). As of 2012, successive records for the smallest June snow cover extent have been set each year in Eurasia since 2008, and in 3 of the past 5 years in North America (Derksen and Brown, 2012).

7.1.2 Altered land cover

Alterations in land cover began thousands of years ago when people began to grow crops, domesticate grazing animals, or even use fire to alter vegetation communities in favor of plants preferred by animals that people hunted. Primary types of land-cover alterations include deforestation, afforestation, grazing, crops, urbanization, mining, wetland drainage, and, more locally, commercial recreational development.

Deforestation

People have reduced global forest cover to about half of its maximum extent during the Holocene and have eliminated most old-growth forests (Montgomery et al., 2003a). Forestry (wood harvest, with anticipated regrowth of the forest) is now practiced in many industrialized countries, but historical deforestation was seldom conducted in a manner conducive to forest regrowth, and minimal attention is currently paid to regrowth in many areas experiencing deforestation in developing countries. Consequently, the discussion that follows here applies primarily to clearcutting, or complete removal of forest cover over areas of varying spatial extent, rather than to selective cutting of a limited number of trees within a forest.

An extensive literature documents the effects of deforestation on rivers from mountains to lowlands and from the boreal regions to the tropics. Thorough reviews are provided by Foley et al. (2005), Scanlon et al. (2007), Douglas (2009), and Wohl (2010). Cutting of trees and the associated building of roads typically greatly increase hillslope sediment yield over a period of a decade or less, and increase water yield over periods of multiple decades until vegetation recovers. Deforestation in the tropics, however, can decrease water yield because of reduced transpiration and precipitation (Costa, 2005; Wohl et al., 2012a). At present, the three main humid tropical forest regions of South America, Africa, and southeast Asia have particularly accelerated and widespread deforestation (Drigo, 2005).

Sediment yields increase as mineral soils exposed and compacted during deforestation become more susceptible to erosion via overland flow, rilling, and landslides and debris flows. Bank erosion, headward expansion of channels, and windthrow of remaining trees can also increase sediment yields. Roads decrease hillslope stability by redistributing weight, changing surface angles, reducing infiltration, altering conveyance via culverts under roads, and initiating gullies (Figure 7.2). Unpaved roads continue to contribute increased fine sediment yields to channels long after harvested vegetation has regrown (Jones et al., 2000). Water yields increase as removal of vegetation reduces interception and transpiration, and compaction decreases infiltration. Changes in sediment and water yield cause changes in streamflow, stream chemistry, and channel morphology. Depending on the magnitude and timing of alterations in sediment and water yields, numerous channel changes can result indirectly from timber harvest and road building (Figure 7.3). Changes in channel morphology can persist for at least several decades (Madej and Ozaki, 2009). Associated processes such as recruitment of instream wood may require two centuries to return to pre-cutting levels because trees must grow to maturity and

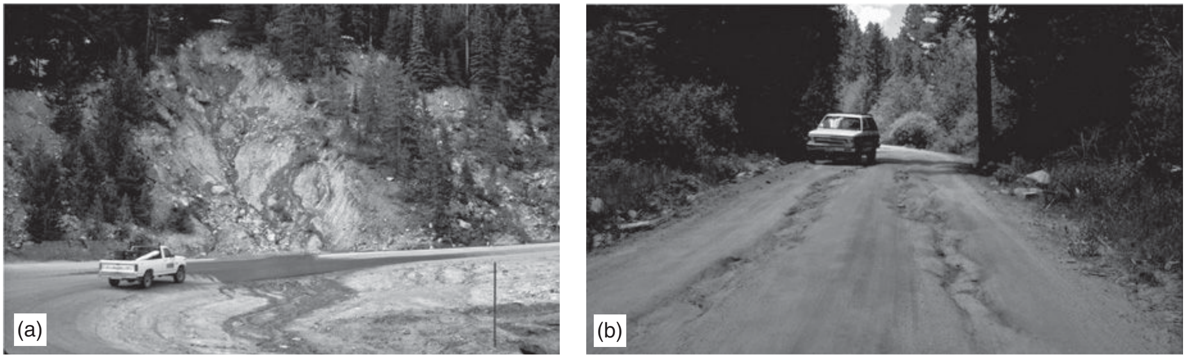

Figure 7.2 Two of the ways in which roads influence sediment yields. (a) In this view of a paved road, the steep face cut into the hillslope above the road bed has destabilized the slope and caused mass movement. (b) Small gullies forming in this unpaved road enhance sediment yield to adjacent streams.

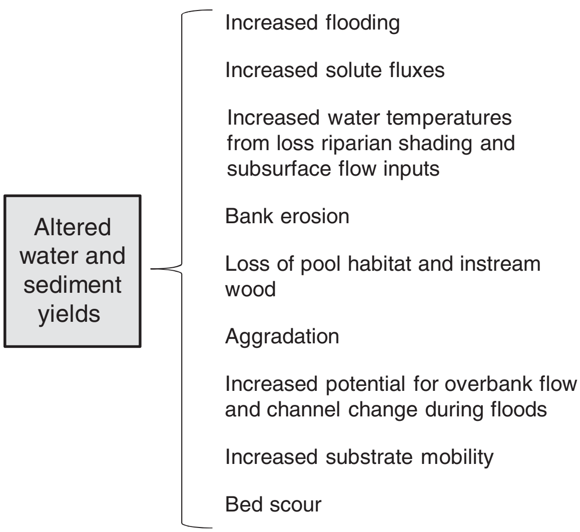

Figure 7.3 Potential channel responses to timber harvest and associated road building as a result of changes in water and sediment yield entering the river network.

then be recruited into and retained within channels (Bragg et al., 2000).

The degree to which water and sediment yields, and thus rivers, are altered depends on

- the portion and spatial distribution of deforested areas within a drainage basin—logging at lower elevations in catchments of mountainous, snowmelt-dominated catchments in southwestern Canada caused little or no change in peak flow because of relatively small low-elevation snowpacks, for example, whereas logging in the upper elevations of these catchments resulted in substantial increases in peak flow (Whitaker et al., 2002);
- the methods used to harvest trees—clearcutting typically results in substantially larger effects on water and sediment yields than selective logging, as shown, for example, by Kasran (1988) in Peninsular Malaysia;
- site characteristics including climate, topography, and soils—subcatchments most responsive to changes in water and sediment yield within the Drôme River basin in France, for example, were high-energy environments of high elevation, high relief ratio, and abundant sediment sources (Liébault et al., 2002); and
- the connectivity between deforested areas and channels—this can be strongly influenced by the extent and placement of roads, which are major sources of sediment and can provide effective conduits for water and sediment from deforested areas to adjacent channels, as shown by gully development from road outlets to streams in southeastern Australia (Croke and Mockler, 2001).

Clearcutting an entire drainage causes major changes in water and sediment yield and rivers. Selective cutting, with riparian buffer strips of unaltered vegetation left in place, minimizes change.

Afforestation

Afforestation, or the regrowth of forest cover, whether natural or human induced, can reverse some of the trends created by deforestation. Documented effects of afforestation include increased infiltration and base flow, decreased runoff and sediment yield, and channel narrowing and deepening. The magnitude and timing of these changes can be complicated by continuing remobilization of sediment eroded during deforestation and stored in colluvial and alluvial features within the drainage basin (Larsen and Román, 2001). If afforestation occurs as a result of declines in population in regions that have been extensively deforested for centuries, as is now occurring in portions of the European Alps, channels that had become stable under former land use patterns can become unstable under newly changing water and sediment yields (Latocha and Migoń, 2006).

Grazing

Upland grazing can have many of the same effects as timber harvest. These include reduced and altered vegetation cover, soil compaction, reduced infiltration, increased runoff and sediment yield, and associated changes in streamflow, channel morphology, and channel stability. The more intense and widespread the grazing, the more severe are these effects. Upland grazing can be practiced with minimal effect if the density of grazing animals is kept below a level that negatively impacts vegetation type and density, and does not compact soils. Overgrazing characterizes mountain livestock husbandry worldwide (Hamilton and Bruijnzeel, 1997), however, leading to land degradation and altered river process and form.

Crops

As with deforestation and upland grazing, the planting of crops alters soil infiltration capacity, soil exposure and erosion, and thus water and sediment yields to channels. Compacted footpaths and roads around crop lands exacerbate these effects. Land drainage associated with crops can increase or decrease runoff.

The magnitude of alterations in water and sediment yield depends on the type and extent of crops, as well as topographic and soil characteristics at the site. Not all crops are created equal. Replacement of grains with potatoes in nineteenth-century Poland, for example, increased sediment yields and flood peaks to the point that meandering channels became braided (Klimek, 1987).

The most common effect of planting crops is increased sediment yield and associated changes in channel pattern and stability. This effect can be documented in sedimentary records of prehistoric land use (Mei-e and Xianmo, 1994). Compilation of numerous case studies indicates that soil erosion under conventional agriculture exceeds rates of soil production and geological erosion by up to several orders of magnitude, a situation that is clearly unsustainable with respect to soil fertility (Montgomery, 2007). Crops can also increase nutrient fluxes to channels as soil nutrients are lost and excess fertilizers are mobilized from uplands (CENR, 2000; Wang et al., 2004; Boyer et al., 2006).

Urbanization

Wolman's (1967a) classic study of the effects of urbanization on sediment yield and channel response delineated a sequence of changes within a small watershed in the Piedmont region of Maryland, USA (Figure S4.25). Subsequent studies from the tropics to the high latitudes document similar sequences that differ only in the details of magnitude and timing (Chin, 2006; Gurnell et al., 2007; Chin et al., 2013).

Removal of vegetation and leveling or artificially contouring the land surface during the initial phase of construction dramatically elevates sediment yields and causes aggradation and planform changes in receiving channels. Completion of construction stabilizes ground surfaces beneath roads, buildings, and lawns, causing sediment yield to decline to a negligible value, and triggering bed coarsening, incision, and bank erosion in receiving channels. Water yield increases as impervious surface area within the contributing basin increases. Storm sewers rapidly drain urban areas, further decreasing infiltration and conveyance time of water, creating flood peaks of shorter duration and higher magnitude for at least small to moderately sized precipitation inputs, and exacerbating channel erosion. (Where storm sewer drainage is fed into the sanitary sewer system, or into buffer areas planted with vegetation, these effects can be minimized.) The magnitude, extent, and speed of change in channel networks reflect factors such as the percentage of the catchment that becomes impervious, the connectivity and conveyance of impervious surfaces, and the characteristics of the receiving channels (Bledsoe and Watson, 2001).

Urbanization typically also involves direct alteration of channels via channelization, bank protection, and other engineering, as well as substantial changes in water chemistry via contaminants moving with runoff and sediment. Headwater channels are commonly paved or otherwise filled, or diverted into sewer systems, effectively eliminating this portion of a river network.

Upland mining

Upland mining that occurs within a watershed but outside of the channel and floodplain can take the form of hard-rock mining of metals or surface or subsurface excavation of coal, construction aggregate, or building stone. Extensive surface disruption or creation of large tailings piles can increase sediment yield to channels (Harden, 2006). Alteration of runoff pathways can increase peak runoff

Figure 7.4 Headwater channels in West Virginia, USA, a region with extensive coal mining and mountain-top removal. (a) A stream in an unmined catchment typically has intermittent or ephemeral flow. (b) A channel downstream from a mining area with valley fill is more likely to have perennial flow, and the increased hydraulic forces result in channel widening, coarsening, and steepening. The channel boundaries become less complex, habitat diversity declines, and disturbances in the form of large flows and bed mobility become more common. (c) Constructed, terraced slopes and "channels" lined with riprap in a mined headwater area. Photographs courtesy of Kristin Jaeger.

(McCormick et al., 2009). Treatment of ores or concentration of naturally occurring toxic contaminants can substantially alter the chemistry of surface and subsurface waters and fluvial sediments (Stoughton and Marcus, 2000). Extreme examples of upland mining impacts come from the practice of mountain-top removal in coal-bearing regions of the eastern United States (Palmer et al., 2010a), in which material overlying coal seams is removed to vertical thicknesses up to 300 m and dumped into adjacent headwater valleys, obliterating surface flow and valley topography (Figure 7.4).

Land drainage

Land drainage has been undertaken for centuries in many parts of the world in order to decrease the extent of standing water at the surface, lower the water table, and improve access to low-lying lands for agriculture and settlement. The English, working with Dutch engineers, had drained more than 38,000 ha in the eastern counties of England (known as the Fens) by 1649 (Simco et al., 2010). George Washington's library included a copy of "Practical Treatise on Draining Bogs and Swampy Ground," first printed in 1775 (Stephens and Stephens, 2006). Extensive wetland drainage occurred in southern Europe during the seventeenth and eighteenth centuries in an attempt to control malaria, which was then believed to result from "miasma," or the moist air coming from marshes and swamps (Davis, 2006). Just as most people today cannot really imagine how much wood was historically in rivers and how many rivers had a multi-thread planform, we have trouble understanding how extensive various types of wetlands—marshes, swamps, fens, and mires—once were, and how much their drainage and loss has altered water and sediment dynamics across large regions (Vileisis, 1997).

Field drainage systems are designed to control surface water and the water table via surface features such as bedded systems used on flat lands growing rice or graded systems used in sloping lands for other crops. Field drainage can also use subsurface features such as horizontal or slightly sloping channels or trenches, wells with pumps, and buried pipe drains. In many cases, an extensive network of collector and main drains exists to transfer water to a gravity outlet structure or a pumping station. Drainage can result in soil compaction and increased runoff, soil erosion, and suspended sediment transport, although the changes are not necessarily substantial (Walling et al., 2003).

Commercial recreational property development

Commercial ski resorts and other spatially extensive changes in land cover undertaken for recreational purposes, such as golf courses, can involve deforestation, road construction, alteration of topography, and transfer and application of large volumes of water and/or pesticides. Such activities can alter water, sediment, and nutrient and contaminant yields to nearby channels, resulting in channel change (Keller et al., 2004; David et al., 2009). Very few studies, however, have systematically evaluated the effects of commercial recreational property development on river networks.

An important consideration in reviewing the effects of changes in land cover on river process and form is that very few locations have had only one change in land cover through time. Most river networks have experienced multiple changes that overlapped in time and space. This complicated history, combined with factors such as equifinality, thresholds, lag times, and complex response, can make it very difficult to decipher cause and effect in relation to any particular past land-cover alteration, or to predict responses to ongoing or likely future alterations in land cover.

A common research technique is to compare otherwise similar rivers with and without a particular land-cover change in order to detect the influence of the land-cover change on river process and form. The river without human impacts reflects reference conditions (Section 7.3.1). This *paired watershed approach* is effective to the extent that other potential control variables for river process and form can be held constant. Where the absence of unaltered drainage basins makes such comparisons infeasible, paleoenvironmental records of past changes (Supplemental Section 3.2.1) or numerical modeling of river response to altered water and sediment yield can be used to infer cause and effect. Understanding of process domain or river style (Section 8.2) can also be used to infer reference conditions and the evolutionary trajectory of a river (Fryirs et al., 2012) in the absence of a single, highly similar, reference watershed.

7.2 Direct impacts

7.2.1 Flow regulation

As discussed briefly in Chapter 3, most of the world's rivers are affected to some degree by flow regulation. Dams have been built since circa 2800 BC (Smith, 1971) for diverse purposes, including water supply, flood control, navigation, and hydroelectric power generation. The construction of dams larger than 15 m tall has accelerated substantially since the 1950s (Goldsmith and Hildyard, 1984; Nilsson et al., 2005). Although relatively few large dams are now built in Europe and North America, numerous large dams are being built and are proposed for rivers in Africa, Asia, and South America. Multiple dams are under consideration in the few large river basins not yet extensively regulated, including the Amazon and the Congo (Wohl, 2011a). Although dams are sometimes promoted as an environmentally benign, "clean" source of hydroelectric power, flow regulation nearly always substantially disrupts physical process and form, water chemistry, and ecological communities along rivers.

As reviewed by Petts and Gurnell (2005), geomorphologists began to pay increasing attention to the effects of dams in the late 1960s and into the 1970s (Wolman, 1967b; Gregory and Park, 1974; Petts, 1979). Coupled with advances in measurement techniques and process-based studies, this led to a series of influential papers during the 1980s on the geomorphic effects of dams on river networks (Petts, 1984; Williams and Wolman, 1984; Carling, 1988).

The specific effects of flow regulation depend on the character of the alterations in water and sediment discharge and on the characteristics of the channel (Grant et al., 2003; Salant et al., 2006; Do Carmo, 2007; Magilligan et al., 2013). The presence of a dam typically

- reduces the mean and the coefficient of variation of annual peak flow—analysis of 29 dams in the central and western United States, for example, revealed decreases in average annual peak flows downstream from the dam that ranged from 3% to 90% (Williams and Wolman, 1984);
- increases minimum flows, as illustrated by portions of the Tennessee and Columbia Rivers in the United States, where annual 7-day low flows have nearly doubled downstream from dams with substantial amounts of reservoir storage (Hirsch et al., 1990;

- shifts the seasonal flow variability—the Colorado River below Glen Canyon and Hoover Dams in Arizona, USA, for example, shifted from a pre-regulation May and June snowmelt peak flow to a broad peak from April to September that corresponds to the period of maximum irrigation and municipal water-supply demands (Hirsch et al., 1990); and
- can greatly increase diurnal flow fluctuations if the dam is operated for hydroelectric power generation, as illustrated by the Connecticut River below Wilder Dam in the northeastern United States, where the number of hydrograph reversals per year have increased by 30% (Magilligan and Nislow, 2001).

Numerous studies document the response of the downstream channel to these changes in hydrology, and to the changes in sediment supply as the majority of bedload and, in some cases suspended load, is trapped upstream from the dam. Geomorphic response reflects the ratio of sediment supply below the dam to that above the dam, as well as the fractional change in frequency of flows transporting sediment (Grant et al., 2003) (Figure 7.5). Where sediment supply decreases substantially and flow competence remains sufficiently high, bed coarsening, channel incision, and bank erosion occur, and these effects can extend hundreds of kilometers downstream from a large dam, as shown for several large dams in North America (Galay, 1983) and for the Changjiang (Yangtze) in China (Xu, 1996). Alternatively, the reduction in peak flows and flood scouring can result in bed fining, aggradation, and narrowing, especially if large sediment sources such as tributary inputs are present just downstream from the dam or encroaching riparian vegetation helps to stabilize sediment, as shown for the West River in Vermont, USA (Curtis et al., 2010). Channel adjustment varies with distance downstream from a dam, but many rivers have sequential dams that cause substantial cumulative disruption of water and sediment fluxes.

Where dams have been present for many decades and discharge measurements are sparse, quantifying the changes in flow regime associated with a dam

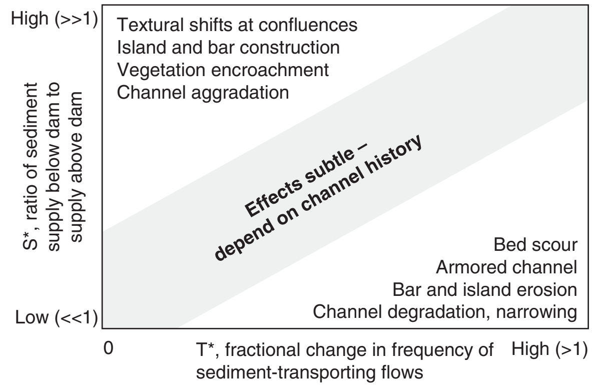

Figure 7.5 Response domain for channel adjustments predicted in response to the presence of a dam. Adjustments are conceptualized in relation to the fractional change in frequency of sediment transporting flows (T*) and the ratio of sediment supply downstream from the dam to supply upstream from the dam (S*). End-member textural and morphologic adjustments are shown. Response of rivers plotting within the shaded diagonal region is likely to be strongly influenced by geological factors, including the history of floods and landslides that leave legacies of large volumes of coarse material and bedrock incision in the valley and channel bottom. (From Grant et al., 2003, Figure 3, p. 209.)

can be difficult. Methods used to quantify hydrologic changes include:

- the range of variability, also known as the indices of hydrologic alteration (IHA) (Supplemental Section 3.2.7), in which the interannual variability of 67 streamflow parameters that reflect magnitude, timing, frequency, duration, and rate of change of discharge are compared pre- and post-regulation (Richter et al., 1996);
- wavelet transform, a mathematical tool used to extract dominant modes of variation from statistically nonstationary signals (Zolezzi et al., 2009); and
- comparison of the seasonal probability density function of natural and regulated streamflows (Botter et al., 2010).

 Of these techniques, IHA is the most commonly used.

In addition to altering water and sediment flux and resultant changes in channel form, dams alter water temperature and chemistry, and the movement of nutrients, plant propagules, instream wood, and organisms. Numerous studies highlight the effects of dams on fish populations (Brooker, 1981; Ligon et al., 1995; Bunn and Arthington, 2002). These effects include physical blockage of migration. Fish lose floodplain spawning and nursery habitat when reduced peak flows limit overbank flooding. Reductions in minimum flow can prevent spawning by restricting access to spawning areas. Fine sediment deposition on gravels used for spawning limits survival of eggs and embryos. Rapid submergence and exposure through dewatering of the varial zone—the shallow borders of channels—can expose and kill fish eggs and embryos. And, disruption of seasonal thermal cues that regulate timing of lifecycles can stress or eliminate fish populations.

Riparian communities also experience numerous disruptions where dams are present (Nilsson et al., 1997; Nilsson and Berggren, 2000; Katz et al., 2005; Merritt and Wohl, 2006). Altered channel morphology or lateral disconnection of the channel and floodplain as a result of reduced peak flows decreases or eliminates freshly scoured surfaces necessary for seedling establishment. Seedlings can be killed by prolonged submersion because of increased base flows. Changes in timing and magnitude of flow relative to release of riparian seeds can disrupt *hydrochory*—the downstream transport of seeds and other plant propagules to germination sites by streamflow. Changing grain-size distribution and moisture content along channel banks and bars can limit seedling establishment. Changes in the riparian water table can cause soil salinization that kills riparian vegetation. The integrated effects of these diverse changes are illustrated by a comparison of four free-flowing and four regulated rivers in northern Sweden (Jansson et al., 2000). The number of plant species and their cover per unit area were lower along regulated rivers, and wind-dispersed (as opposed to water-dispersed) species were more common along regulated rivers.

Global and regional syntheses indicate that aquatic and riparian species are becoming increasingly homogeneous—a few hardy generalist species tend to dominate many communities—as a result of flow regulation (Dynesius and Nilsson, 1994; Moyle and Mount, 2007; Poff et al., 2007; Braatne et al., 2008). A common theme among diverse case studies of regulated rivers is the loss of physical and ecological complexity (Surian, 1999; Graf et al., 2002). This can be conceptualized with changes in hydrology and sediment supply as first-order effects; changes in hydraulics, substrate mobility, and channel form as second-order effects; and changes in biota as third-order effects (Burke et al., 2009) (Figure 7.6).

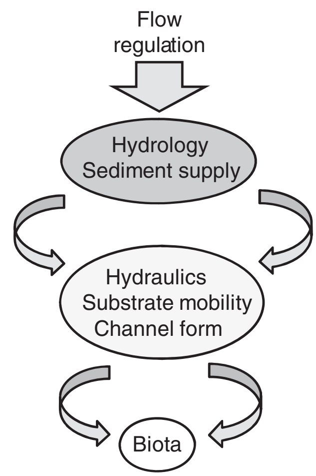

Figure 7.6 Schematic illustration of the changes in a river as a result of flow regulation.

Dams began to be removed in some portions of the United States at the end of the twentieth century. These were mostly relatively small, nineteenth- century dams built for a purpose that no longer existed, such as powering a local mill that was long gone. A primary concern with dam removal is remobilization of sediment stored behind the dam, particularly if this sediment contains toxic contaminants or high nutrient concentrations (Hart et al., 2002; Pizzuto, 2002). Issues of increased mobility for aquatic organisms and loss of lentic (still water) habitat can also be important (Stanley and Doyle, 2003).

Removal of large dams is typically not economically or politically feasible, but disruptions to river ecosystems can be reduced by modifying the operations of the dam (Graf, 2001). The most widely studied example involves experimental flood releases from Glen Canyon Dam on the Colorado River just upstream from Grand Canyon National Park in the United States (Collier et al., 1997; Wright et al., 2008). These releases are timed in order to achieve mobilization and downstream redistribution of sediment inputs from tributaries just downstream from Glen Canyon Dam. The objective of redistributing sediment is to enhance channel-margin sand deposits and associated camping sites and channel-margin fish habitat.

The complexity of most river ecosystems, and the complications introduced by other factors such as exotic species, makes it extraordinarily difficult to predict and create targeted effects on these ecosystems by modifying dam operations, particularly when each experimental flow release is expensive and comes under intense public scrutiny. Nonetheless, the adaptive management being used on the Colorado River, in which hypotheses regarding river response to experimental flow releases are tested, and modified hypotheses are used to design the next experimental release (Wohl et al., 2008), is now being applied to other regulated rivers (Bednarek and Hart, 2005; Jorde et al., 2008).

Flow regulation via diversion of flow from a source to a receiving channel has occurred primarily in dry regions and on relatively small channels, although diversions are present in a range of environments from northern Canada (Kellerhals et al., 1979) to southern Australia. Massive flow diversions are now being proposed or built for some of the world's largest rivers, however, including the South–North Transfer Water Project to transfer water from China's Huang He (Yellow River) north to the Changjiang (Yangtze River); the Jonglei Canal Project that would divert waters more than 300 km from the Sudd wetlands along the Nile; and the Solomon pipeline that would divert water 1000 km south from the mouth of the Congo River to Namibia (Wohl, 2011a).

Diversion has most commonly been undertaken for water supply or flood control. Diversion tends to reduce base flow and flood peaks on the source channel, and increase these parameters in the receiving channel. As with dams, changes in flow regime associated with diversions disrupt sediment dynamics, channel morphology, and aquatic riparian communities in both source and receiving streams. The magnitude of these disruptions depends on the magnitude of hydrologic changes and the characteristics of the affected channels (Ryan, 1997; Parker et al., 2003; Wohl and Dust, 2012) (Figure S7.1).

Among the most insidious ecological effects of flow diversions are the introductions of new species from one river network to another. Millions of dollars are now being spent in the United States, for example, to construct barriers using concrete, wire mesh, and electrical fields designed to prevent four introduced species of Asian carp from migrating through a canal connecting the headwaters of the Illinois River system, where the fish are established, into the Great Lakes ecosystem via a canal between the Illinois and Lake Michigan (Sandiford, 2009). In southeastern Australia, water impounded and diverted from the headwaters of the eastward-draining Snowy River into the headwaters of the westward-draining Murray River catchment since 1967 has been accompanied by pioneering fish species, including climbing galaxias (*Galaxias brevipinnis*), which may compete with many of the already endangered native species of the Murray–Darling basin (Waters et al., 2002).

7.2.2 Altered channel form and connectivity

In addition to regulating flow, humans have been directly altering channel form for centuries. The intent behind these alterations is as diverse as the alterations:

- increasing channel conveyance by dredging the channel or building levees to reduce overbank flooding;
- straightening sinuous channels and stabilizing banks to reduce channel mobility and bank erosion;
- removing instream wood or beaver dams to increase conveyance and downstream water supplies, to reduce overbank flooding, or to enhance fish passage;

- building check dams and weirs to limit bed incision and/or downstream movement of coarse bedload;
- extending channel networks via canals that can connect the headwaters of two networks that share a common drainage divide, or can shift water across tens to hundreds of kilometers to a different river network;
- burying or laterally shifting channels perceived as inconveniently placed with respect to urban areas, transportation corridors, or other land uses;
- removing sediment from channels to mine placer metals or construction aggregate; and
- reconfiguring a channel to a form considered more esthetically pleasing.

Early, well-documented examples include Li Ping's extensive system of irrigation canals and flood-control structures built more than 2100 years ago in the Szechuan region of China (Yao, 1943). Levees were constructed along the Yodo River in Japan during the fourth century AD. Check dams were built along mountain rivers in Japan at least as early as 806 AD. (Japanese Ministry of Construction, 1993). Wood has been cleared from channels since the twelfth century in France (Piégay and Gurnell, 1997). Flood embankments were built along Italy's Po River during the fourteenth century (Braga and Gervasoni, 1989). In North and South America, Australia, New Zealand, and other areas colonized by Europeans, alterations to channel form typically start soon after European settlement.

As with flow regulation, an extensive literature documents numerous case studies of altered channel form and associated changes in longitudinal, lateral, and vertical connectivity. This section provides a brief review of some of the most widespread types of direct channel alteration.

Levees are as ubiquitous along lowland (as opposed to mountain) rivers as are dams along all rivers. And, like dams, levees have a very long history. Linear mounds constructed along rivers to limit overbank flooding are known as levees, dikes or dykes, embankments, and floodbanks, among other things (Petroski, 2006). Levees can be permanent or temporary, but they have been constructed at least since 2600 BC, when levees were built in the Indus River valley. Levees date to more than 3000 years ago in Egypt along the Nile, and extensive levees were built by ancient Mesopotamian civilizations and the Chinese. At present, particularly extensive permanent levees line the Danube, Po, Rhine, Meuse, and Rhone Rivers of Europe, the Mississippi and Sacramento River systems in the United States, and most of the large rivers of China. The Mississippi levee system is one of the world's largest. Begun by French settlers in Louisiana during the eighteenth century, it now includes more than 5600 km of levees along the middle and lower portions of the river, and most major tributaries such as the Illinois, Ohio, and Missouri Rivers also have extensive levees.

Levees reduce or eliminate all of the channel–floodplain exchanges discussed in Chapter 6. Levees facilitate higher magnitude, shorter duration floods, and exacerbate flooding in downstream areas without levees. As levees cause flood peak discharge and shear stress to increase, the river bed between the levees coarsens and may incise (Frings et al., 2009). Levees severely reduce lateral connectivity of water, sediment, organic matter, and organisms, leading to loss of habitat, animal abundance, and biodiversity in channel and floodplain environments (Hohensinner et al., 2004). The floodplain can be a major sediment source as the result of bank erosion along sinuous rivers (this is a net sediment source only if the eroded sediment added to the channel is not compensated by deposition on the opposing bar surface), but levees and associated bank protection truncate this sediment exchange, commonly leading to channel incision (Kesel, 2003).

Along rivers with extremely high suspended sediment loads, including the Huang He in China, levees limit lateral channel movement and facilitate sediment deposition within the channel. With time, these channels have become elevated above the surrounding floodplain, a situation that the Chinese describe as "hanging rivers" (Xu, 2004). Levees have been built along the lower Huang He since 475 BC, and the bed of the river is now mostly 3–5 m above the floodplain beyond the levees. In some places the river bed is 10 m above the surroundings. Water seeps into the ground from this elevated river at rates

Figure 7.7 A highly stabilized urban river: the Vienna River in Vienna, Austria.

of 1.29 m^3 per meter of river length per day during high flow and 0.59 m^3 during low flow, reducing discharge by an estimated 83 million m^3 each year (Xu, 2004). Along with other changes in the drainage and with consumptive water use, this seepage loss caused the lower Huang He to become ephemeral starting in 1972, a condition for which there was no earlier evidence in the historical record. With time, periods of no flow in this large river have become more frequent and the point of no flow has moved farther upstream (Xu, 2004).

Dredging, channelization (sometimes known as canalization in Europe), and bank stabilization commonly occur together and in association with levees, the intent being to make the channel more physically uniform in dimensions, including flow depth, and to limit overbank flooding, bank erosion, and lateral channel movements. *Dredging* involves physically removing sediment from the channel in order to increase cross-sectional area and flow conveyance, usually for purposes of navigation or flood control. *Channelization* involves making the channel straighter and larger in cross-sectional area, typically via dredging and other forms of sediment removal such as digging out the streambanks with heavy machinery. *Bank stabilization* (also known as *bank hardening*) involves increasing the erosional resistance of the streambanks using methods that range from planting riparian vegetation, to covering the banks in large boulders (riprap) or concrete (Figure 7.7).

Collectively, dredging, channelization, and bank stabilization are so ubiquitous and of such antiquity in many regions that most people have no idea what altered rivers looked like historically. Channels at opposite ends of the drainage basin in the Danube River of Europe provide an example.

Alpine headwater tributaries of the Danube have been extensively "trained" since the sixteenth century. *Training* refers to engineering channel form and mobility, and includes bank stabilization, check dams, and confinement of braided channels into a single, straightened, and stabilized channel (Figure 7.8). Larger, mid-basin tributaries have been similarly engineered since the nineteenth century to alter braided channels to single channels. Sediment introduced to channels from hillslopes and formerly transported downstream has been stored in check dams and sediment detention basins in the headwaters. Sediment temporarily stored in floodplains and

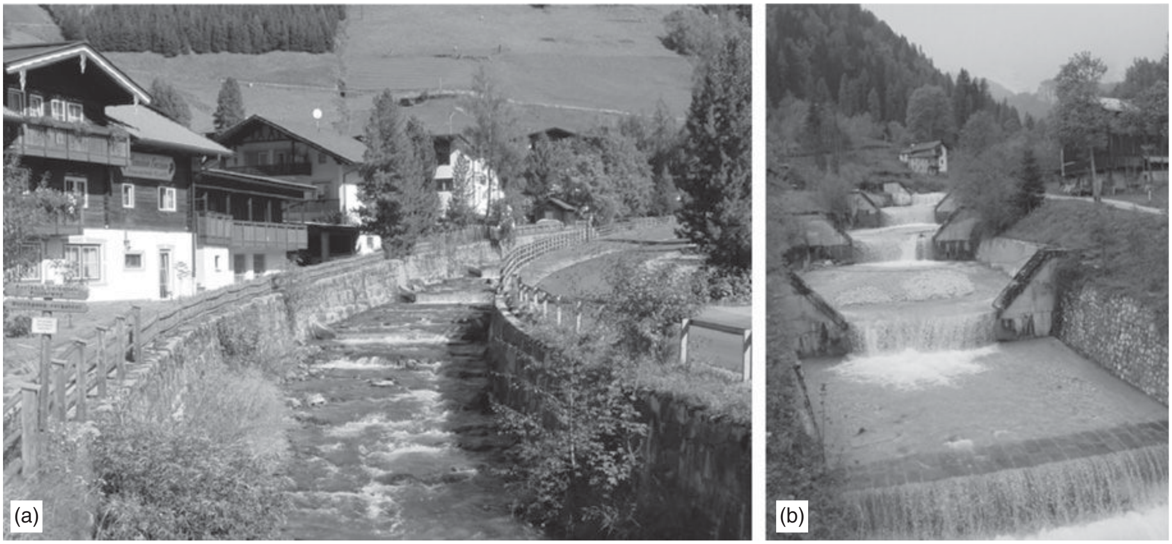

Figure 7.8 Examples of trained alpine rivers. (a) In this view upstream in the town of St. Jakob, Austria, the river is channelized, with stabilized banks and regularly spaced concrete steps. (b) This river in the Italian Dolomites flows between stabilized banks and over pronounced artificial steps.

alluvial fans along mid-network valley bottoms has been stabilized to limit mobilization during floods. The net effect of these alterations has been to substantially reduce sediment supply to downstream river segments, exacerbating incision and bank erosion (Habersack and Piégay, 2008).

Floodplain habitat has also been reduced. The Upper Danube drainage, for example, has lost 95% of historically present floodplain habitat (Bloesch, 2003). Lateral connectivity between the main channel and secondary channels has declined, along with associated river complexity, aquatic habitat and productivity (Hohensinner et al., 2004), and abundance and diversity of fish (Aarts et al., 2004).

Downstream from the Alps, the Danube alternately flows through narrow canyons and broad alluvial basins that now contain cities such as Vienna, Austria and Bratislava, Slovakia. Historically, the Danube had a braided or anastomosing planform in these basins, with multiple channels separated by side arms, backwaters, and forested floodplain. River training began along these segments of the river in the eighteenth century, and accelerated during the nineteenth century, resulting in a single, straight, highly stabilized channel relative to historical, multi-thread channel planforms (Pišút, 2002). As in the headwater tributaries, these alterations have resulted in diverse problems from bed and bank erosion that undermine engineering infrastructure along the stabilized channel to dramatic losses in abundance and diversity of aquatic and riparian species (Bloesch, 2003).

Channelization of lowland alluvial channels, typically undertaken to increase conveyance and reduce overbank flooding, was particularly widespread in portions of the United States until the 1970s (Wohl, 2004a). By making channels straighter, steeper, and less complex, channelization triggered responses that varied from incision and widening in the channelized reaches and upstream segments of the drainage to aggradation and exacerbated overbank flooding in downstream reaches (Schoof, 1980; Simon, 1994; Wyżga, 2001). Channel evolution models (Section 5.1.6) were developed to describe the successive responses through time of channelized streams.

Check dams are a type of channel alteration common in steep, mountain channels (Lenzi, 2002), and are primarily designed to retain sediment. Check dams share some similarities with weirs. *Weirs* are designed to provide grade control, or local base level, in lowland alluvial channels that are actively incising, sometimes because they have been artificially straightened or channelized. Check dams are also used to create a local base level that limits upstream migration of headcuts in steep channels. By storing sediment, check dams are designed to limit downstream deposition, overbank flooding, and channel avulsion, and to enhance profile irregularity and dissipation of flow energy. Check dams are particularly abundant in steep channels of Europe and Asia (Castillo et al., 2007; Shieh et al., 2007).

Closed check dams built of concrete or rock are designed to trap all sediments until the structure fills, at which point the sediment can be removed or another structure can be built (Figure S7.2). *Open check dams* built of some type of very coarse mesh can pass finer sediment downstream (Figure S7.3). Check dams are typically built in series, rather than singly, along a channel.

Although check dams typically do not strongly affect flow regime, they greatly disrupt sediment fluxes. Disruption of sediment fluxes commonly causes exacerbated channel erosion downstream (Wyżga, 1991), as well as altering channel geometry, substrate characteristics, and aquatic and riparian communities (Bombino et al., 2009). New approaches to check dams include designing structures that mimic naturally occurring step–pool bedforms both in appearance and function (Lenzi, 2002).

Although people have built hundreds of thousands of dams and check dams along rivers throughout the world, they have also actively removed naturally occurring channel-spanning obstructions, including instream wood and beaver dams (Supplemental Section 3.2.6). Beavers, once extremely abundant and widespread throughout Europe and North America, are now highly restricted in distribution and abundance, and are effectively absent from much of their former range. Loss of beaver populations results in loss of beaver dams. This causes increases in conveyance, flow velocity, sediment erosion and transport, as well as decreases in overbank flooding, riparian water tables, nutrient storage, and aquatic and riparian habitat diversity and abundance (Naiman et al., 1988; Westbrook et al., 2006; Burchsted et al., 2010). (Conversely, beavers introduced to Chile in 1946 have locally eliminated riparian forests and greatly expanded meadow environments (Anderson et al., 2006b).)

Wood has been removed from channels for centuries to reduce flooding, to enhance river navigation and fish passage, and because it is considered esthetically unattractive (Section 5.6.2). Wood removal has gone on so long that most people have no understanding of how abundant and widespread instream wood was historically, and consequently harbor negative attitudes toward wood (Chin et al., 2008). Records of wood removal indicate that individual pieces, logjams and, in very large rivers, log rafts extending many kilometers along a river, were present from the smallest headwater channels to the largest alluvial rivers anywhere in the boreal and temperate zones where forest was historically present (Sedell and Froggatt, 1984; Piégay and Gurnell, 1997; Phillips and Park, 2009). (Limited studies of wood in tropical environments indicate that, although instream wood performs important geomorphic and ecological functions in tropical streams, the wood tends to be much more transient as a result of greater discharge per unit drainage area and faster decay rates (Wohl et al., 2012b).)

Instream wood provides hydraulic resistance, enhances storage of sediment and organic matter, increases substrate diversity and aquatic habitat, and enhances overbank flooding and channel–floodplain connectivity (Section 5.6.2). Active removal of instream wood, and passive loss of such wood because of reduced recruitment from riparian zones and hillslopes, has caused reduced stability and complexity along a wide variety of channels. Attempts to actively reintroduce instream wood in the form of engineered logjams (Southerland and Reckendorf, 2010; Abbe and Brooks, 2011) or individual pieces remain limited by lack of detailed, quantitative understanding of how wood characteristics (abundance, size, spatial distribution along a channel or river network) relate to channel form and process.

Another form of channel alteration involves actively or passively altering riparian vegetation.

Active alteration includes removal and replanting. Passive alteration includes changes associated with riparian grazing or the spread of invasive, exotic species. Removing or replanting riparian vegetation alters bank erodibility, as well as near- and overbank flow resistance and sediment deposition (Section 4.5).

Riparian grazing affects the density, type, and spatial extent of riparian vegetation, and also exposes the bank substrate and directly erodes the banks via trampling (Trimble and Mendel, 1995). Grazing animals can create ramps along stream banks and trails along the floodplain that enhance localized erosion (Cooke and Reeves, 1976). Enhanced bank and overbank erosion can result in deposition of fine sediments on the bed, reducing pool volume, spawning and macroinvertebrate habitat, and hyporheic exchange (Myers and Swanson, 1996). Increased water temperature and reduced dissolved oxygen can result from loss of shading by overhanging and riparian vegetation and contamination by fecal material of grazing animals. Several studies indicate that these effects can be rapidly reduced or eliminated if short-duration grazing or grazing exclosures replace continuous grazing of the riparian corridor (Magilligan and McDowell, 1997; Magner et al., 2008).

Invasive, exotic riparian species with different characteristics than native riparian species can alter the density of riparian plants, and thus influence streambank resistance to erosion and near- and overbank sedimentation (Graf, 1978; Allred and Schmidt, 1999). Exotic plants can also change patterns of water uptake and transpiration, thus altering streamflow and riparian water tables, as well as nutrient cycling and riparian habitat for other species (Hultine and Bush, 2011). Invasive, exotic riparian species are particularly widespread and well documented in southeastern Australia and the western United States.

Although *instream mining* for sand and gravel used in construction and for placer deposits of precious metals has occurred for millennia in some regions, systematic studies of the physical and ecological effects of these activities began only during the twentieth century (Gilbert, 1917). By altering substrate grain-size distribution and abundance, all forms of instream mining disrupt sediment dynamics and cause a variety of channel adjustments. Large, localized excavations can initiate a knickpoint (Wishart et al., 2008). Bed and bank erosion downstream from mining can be exacerbated by reduced sediment supply (Lagasse et al., 1980). Sediment deficit at the mining site can result in bed coarsening and loss of aquatic habitat (Kondolf, 1997). Disruption of a coarse surface layer can increase sediment mobility and downstream turbidity, aggradation, overbank deposition, and lateral channel mobility (James, 1997; Parker et al., 1997) (Figure 7.9). These changes can be so severe that they cause a meandering channel to become braided, as in the case of the Middle Fork of the South Platte River in Colorado, USA (Hilmes and Wohl, 1995). Knighton (1989) documented increases in channel width of up to 300% following alluvial tin mining on the Ringarooma River in Australia, and the development of a braided planform that later reverted to a single channel when mining ceased and sediment supply declined.

Mining-related changes in channel form and process can significantly disrupt aquatic communities, from primary production by algae to survival of fish. This was documented for the headwaters of the Chatanika River in Alaska by Van Nieuwenhuyse and LaPerriere (1986). Comparing otherwise analogous mined and unmined headwater tributaries, they found that the watersheds with mining had turbidity values up to two orders of magnitude higher, which resulted in 50% reduction in primary productivity in a moderately mined stream, and no detectable primary production in a heavily mined stream.

Diversion of flow to process placer deposits further disrupts river form and process, and toxic material such as mercury associated with placer mining can contaminate aquatic and riparian systems for decades to centuries after mining ceases (Taylor and Kesterton, 2002; Macklin et al., 2006). Metal contamination from sulfide-ore mining in the headwaters of Soda Butte Creek continues to contaminate riparian vegetation and aquatic insects along the creek in Yellowstone National Park, USA,

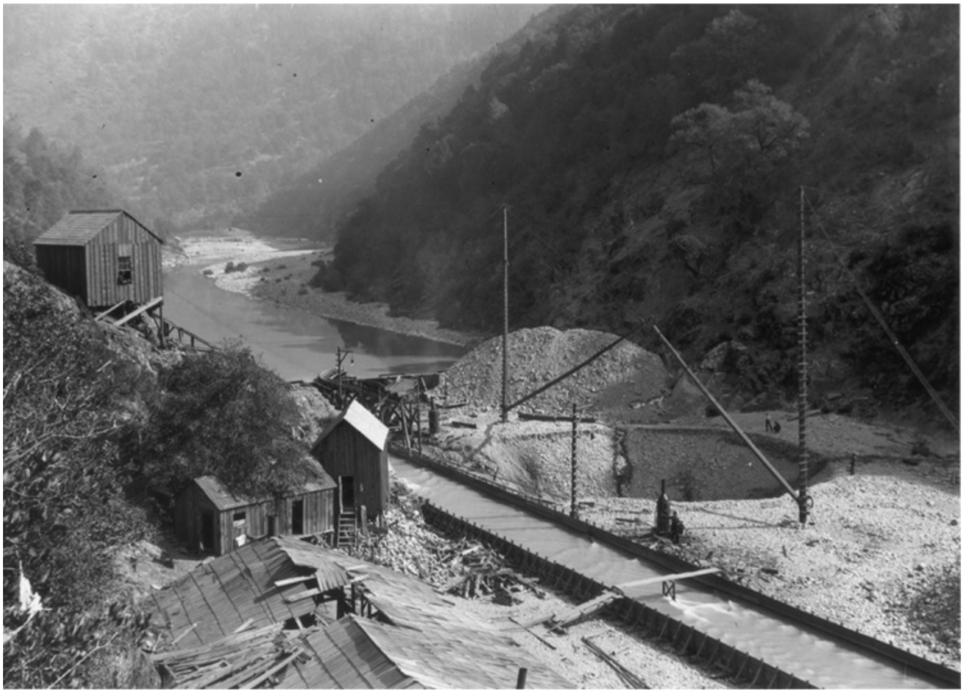

Figure 7.9 Historical photograph of placer mining pit along the Middle Fork of the American River, Monte Rio Mining Company, 1903. (Photograph courtesy of the Bancroft Library, University of California, Berkeley)

for example, following a tailings dam failure in 1950 (Marcus et al., 2001). Mining and associated river contamination in parts of Europe, in particular, began centuries ago (Macklin et al., 2006).

In summary, it is important to once more emphasize that the cumulative, combined effects of direct and indirect human alteration of rivers are ubiquitous (Gregory, 2006; James and Lecce, 2013). River networks in even seemingly remote areas of polar regions, the tropics, vast inland deserts, and sparsely settled mountains have been directly and indirectly altered by a variety of human activities (Wohl, 2006, 2011a; Comiti, 2012). In many cases, these activities and the resulting changes in river systems occurred so long ago that there is little or no historical record, let alone collective awareness, of them.

Increasing scientific attention to prehistoric or historical activities that altered rivers has led to use of the phrases *legacy effects* or *legacy sediments* (Walter and Merritts, 2008). A legacy can be defined as "something received from a predecessor or from the past." In the context of rivers, legacy refers to sediments or channel form resulting from historical land uses. An example comes from a study of streams in the mid-Atlantic Piedmont of the eastern United States, where streams that historically had small, anabranching channels with extensive vegetated wetlands and low sediment accumulation were transformed by the construction of tens of thousands of milldams during the seventeenth to nineteenth centuries (Walter and Merritts, 2008). After dam construction, 1–5 m of fine, slackwater sediment buried the original channel network and wetlands upstream from each dam. As the dams were abandoned, they fell into disrepair and were breached, causing channel incision down to Pleistocene basal gravels. The resulting incised, meandering gravel-bed streams were assumed to represent a channel form with little human influence until Walter and Merritts (2008) undertook detailed historical and stratigraphic reconstructions of the region and revealed the legacy of historical alteration. These insights have proved controversial because of differing views of how to protect and restore river corridors in the region. At present there is a dichotomy between those who seek to protect single-thread channels with riparian forests and those who seek to restore multi-thread channels with floodplain wetlands. Similarly dense networks of mill dams were present throughout Britain and much of northern Europe.

7.3 River management in an environmental context

This portion of the chapter focuses on river management undertaken specifically to restore rivers in an environmental context. People have been managing—or at least attempting to manage—rivers for millennia. Past management actions were typically undertaken with the intent of making rivers or associated resources more conveniently accessible to humans: damming rivers to ensure water supply, for example, or building levees and channelizing rivers to enhance agricultural use or settlement on floodplains. Although river restoration and rehabilitation are sometimes viewed as being fundamentally different than past river management, they are the latest iteration of trying to reconfigure rivers to conform to human expectations: in the case of restoration, expectations of more natural or esthetically attractive rivers.

Restoration activities such as planting riparian vegetation to stabilize streambanks date to the seventeenth century in Europe (Evette et al., 2009), but projects designed to restore rivers have increased dramatically in number and scope since the 1990s (Bernhardt et al., 2005), particularly in the United States, Western Europe, Australia, and New Zealand. As with any form of river management, some projects have largely achieved their original purpose, whereas others have been thorough failures. The rationales that underpin river restoration, and the factors that result in success or failure, are worth examining because the ability to restore rivers provides an effective measure of our understanding of river process and form.

7.3.1 Reference conditions

Any river restoration is undertaken to achieve some desired end result of river process and form. Restoration is undertaken for a variety of reasons, including those related to recreation, water quality, esthetics, protection of aquatic and riparian species, bank stabilization, fish passage, flow modification and dam removal, and creation of a more natural environment (Table 7.2) (Bernhardt et al., 2005). The latter point is perhaps the most difficult, because achieving a more natural environment entails addressing at least one fundamental question: what is natural? (Graf, 1996; Wohl, 2011d).

Natural is commonly assumed to imply minimal human alteration, although natural is increasingly being defined in terms of *geomorphic integrity* (Graf, 2001), or the ability of the river to adjust to existing conditions (Fryirs and Brierley, 2009). Humans have been manipulating natural landscapes for many thousands of years by using fire to alter land cover, selectively hunting some animals to extinction, domesticating plants and animals and then altering ecosystems to favor domesticated species, and cutting trees for fuel and building materials. So at what point in history do we consider a given ecosystem to have last been natural: prior to agriculture, prior to the Industrial Revolution, or prior to some arbitrary human population density?

Whatever (pre)historical period is chosen, the characteristics of rivers during that period are typically known as *reference conditions*, which can also be defined as the best available conditions that could be expected at a site (Norris and Thoms, 1999). The latter definition can be highly problematic, however, because great disagreement or uncertainty can arise as to what constitutes "best available." Reference conditions encompass all of the aspects of river process and form discussed to this point, including flow regime; sediment regime; water chemistry; substrate; bedforms; channel morphology, planform, and gradient; and longitudinal, lateral, and vertical connectivity.

In a region where some river basins have undergone minimal human alteration, contemporary rivers can provide reference conditions for altered

Table 7.2 NRRSS working group list of goal categories and operational definitions for river restoration projects.

Category	Description
Esthetics/recreation/education	Activities that increase community value: use, appearance, access, safety, and knowledge
Bank stabilization	Practices designed to reduce or eliminate erosion of banks
Channel reconfiguration	Alteration of channel geometry: includes restoration of meanders and in-channel structures that alter river thalweg
Dam removal/retrofit	Removal of dams and weirs or modifications to existing dams to reduce negative ecological impacts (excludes dam modifications solely for improving fish passage)
Fish passage	Removal of barriers to longitudinal migration of fishes: includes physical removal of barriers, construction of alternative pathways, and construction of barriers to prevent access by undesirable species
Floodplain reconnection	Practices that increase overbank flows and flux of organisms and materials between channel and floodplain
Flow modification	Practices that alter the timing and delivery of water quantity (does not include stormwater management)
Instream habitat improvement	Altering structural complexity (bedforms, cross-sectional geometry, substrate, hydraulics) to increase habitat availability and diversity for target organisms and provide breeding habitat and refugia from disturbance and predation
Instream species management	Practices that directly alter aquatic native species distribution and abundance through the addition (stocking) or translocation of plant and animal species and/or removal of exotic species
Land acquisition	Practices that obtain lease, title, or easements for streamside land for the explicit purpose of preservation or removal of impacting agents and/or to facilitate future restoration projects
Riparian management	Revegetation of riparian zone and/or removal of exotic species of plants and animals
Stormwater management	Special case of flow modification that includes the construction and management of structures (ponds, wetlands, flow regulators) in urban areas to modify the release of storm runoff
Water quality management	Practices that protect existing water quality or change the chemical composition and/or suspended load, including remediation of acid mine drainage

Source: From Bernhardt et al. (2007), Table 1.

river basins. This approach must be used with caution, however, because contemporary conditions constitute a "snapshot" in time that reflects only a single state or a limited portion of the fluctuations that naturally occur in rivers (SER, 2002).

In many regions of the world there are no relatively unaltered rivers, so reference conditions must be inferred from historical, botanical, and geologic records (Supplemental Section 3.2.1) (Morgan et al., 1994; Nonaka and Spies, 2005; Stoddard et al., 2006; Wohl, 2011d). Lack of information on reference conditions, as well as continuing change in catchment parameters, can limit the usefulness of reference conditions (Hughes et al., 2005; Newson and Large, 2006). Consequently, reference conditions may be most appropriate as an ideal rather than as a goal for restoration (Osborne et al., 1993).

A tremendous amount of effort may be necessary to characterize reference conditions, not least because river process and form can vary substantially across even a relatively small drainage basin, and because rivers are never static in time. Even in the absence of human manipulation, rivers undergo fluctuations in process and form associated with natural events such as floods or droughts, landslides, wildfires, tectonic uplift or subsidence, and continuing adjustment of the river to long-term changes in climate or tectonics. Consequently, a key component of understanding reference conditions is being able to quantify the *natural or historical range of variability* (HRV) for a given parameter or set of parameters (Morgan et al., 1994; Nonaka and Spies, 2005; Wohl, 2011d).

An example of quantifying HRV comes from a segment of the Upper Colorado River in Rocky Mountain National Park in the United States. The national park was designated in 1915. Prior to that, a flow diversion known as the Grand Ditch was excavated across steep slopes uphill from, and parallel to, the Colorado River. The Grand Ditch cuts across a series of small tributaries flowing into the Colorado River from the west, diverting their waters into the ditch and then into the adjacent Poudre River drainage. The unlined ditch periodically overtops during high snowmelt flows, triggering debris flows that enter the Colorado River valley and the national park. A particularly large debris flow in 2003 buried much of a large wetland along the river and resulted in deposition of ~36,000 m^3 of sediment along the river. In a subsequent lawsuit between the national park service and the private company operating the irrigation ditch, two of the primary questions were whether the 2003 debris flow was beyond the natural range of variability for that site, and how best to restore the river and wetland altered by the debris flow. The question regarding range of variability was not so obvious, because the bedrock along both sides of the Upper Colorado River exhibits hydrothermal alteration that creates faster weathering and erosion. Consequently, the upper river experiences more frequent debris flows than other, adjacent regions, including the eastern side of the valley, which is not affected by Grand Ditch. Examination of the stratigraphy in the Upper Colorado River valley, using ground-penetrating radar and augering to obtain sediment samples and material for radiocarbon dating, produced an ~4300-year record of sedimentation (Rubin et al., 2012). This record indicated a marked increase in the volume of debris-flow deposition on the valley bottom subsequent to construction of Grand Ditch, suggesting that hillslope instability and associated sediment yield to the river corridor coincident with the ditch had in fact exceeded the HRV.

Ongoing climate change, abundant and widespread invasive species, and human population growth and resource use cause some scientists and managers to question the relevance of HRV (Safford et al., 2008). If the world already looks fundamentally different than prior to human manipulation, and will grow increasingly different in the future, what do past river process and form matter? Other scientists and managers contend that, even if a river cannot be restored to HRV, detailed, quantitative understanding of prior and existing river characteristics can inform management by constraining the range of potential river process and form.

Knowledge of HRV provides insight into the conditions that native riverine species or communities might require for survival, as well as the thresholds or minimum values of process or form that must be maintained in order to sustain biological communities (e.g., flood threshold for overbank flooding that provides fish access to the floodplain for spawning and nursery habitat). Knowledge of HRV also facilitates delineation of the spatial distribution of different suites of geomorphic processes, such as portions of a mountainous headwater catchment dominated by debris flows versus portions dominated by fluvial processes. This facilitates evaluating the location, relative rarity, and connectivity of sensitive stream segments that are likely to respond to alterations or that contain biologically unique communities (McDonald et al., 2004; Brierley and Fryirs, 2005; Wohl et al., 2007). Insight into HRV provides information on the relative magnitude of variation in specific river attributes among process domains (Wohl, 2011d). Knowledge of the HRV thus underpins our understanding of process and form in any river network.

7.3.2 Restoration

Restoration is undertaken for many reasons, and at many scales, from a single river segment only a few hundred meters in length, to entire large drainage basins. The National River Restoration

Science Synthesis database includes more than 37,000 restoration projects in the United States. Most of the projects in this database were implemented on river segments less than 1 km in length (Bernhardt et al., 2005), although the most high-profile restoration projects are those involving large segments of river basins, such as the Grand Canyon of the Colorado River in Arizona (Melis, 2011) or the Kissimmee River in Florida (Toth et al., 1993; Warne et al., 2000; Wohl et al., 2008), both in the United States, or the Danube River in western Europe (Tockner et al., 1998, 1999; Bloesch and Sieber, 2003). Restoration in the context of this chapter is used to include both *restoration*—a return to a close approximation of the river condition prior to disturbance (however disturbance may be defined), and *rehabilitation*—improvements of a visual nature, sometimes described as "putting the channel back into good condition" (however good condition may be defined) (National Academy, 1992).

Three basic approaches to river restoration are currently employed (Palmer et al., 1997; McDonald et al., 2004). The *"field of dreams" approach* uses traditional engineering techniques to modify river form to a desired condition with the expectation that this will create processes necessary to maintain that form. This is the most widely used approach for restoration on relatively short river segments. This approach is named in reference to an American movie about baseball that became famous for the phrase "build it and they will come." In the case of river restoration, the implication is that restoring river form will also restore function, so that organisms such as fish will return and thrive.

The *"system function" approach* identifies and alters the initial conditions required to achieve restoration goals. This could involve modifying fine sediment inputs to a targeted segment of river by establishing riparian buffer strips, for example, or adding instream wood to increase flow resistance and sediment retention. The underlying rationale is that modifying initial conditions such as water or sediment inputs will cause river process and form to adjust in a desired manner.

The *"keystone" method* identifies and incorporates crucial components of process and form, and recognizes uncertainty in the resulting river responses. Riffle–pool sequences might be the keystones of river form and process in a project designed to restore fish habitat, so that restoration focuses on the parameters necessary to create and maintain riffles and pools.

River restoration as currently practiced is not commonly scientific because hypotheses regarding river response to a given restoration action are not posed and tested. A large-scale survey of river restoration in the United States found that fewer than 10% of projects included any form of monitoring or assessment, although projects with higher costs were more likely to be monitored (Bernhardt et al., 2005). Less than half of all projects set measurable objectives for the project, although nearly two-thirds of project managers felt that projects were "completely successful" (Bernhardt et al., 2007). This reflects the fact that perceived ecological degradation typically motivated the projects. Post-project appearance and positive public opinion were the most commonly used metrics of success (Bernhardt et al., 2007), however, rather than more objective metrics or metrics grounded in scientific understanding of river process and form.

The lack of monitoring for restoration effectiveness is highlighted by a consideration of the history of river restoration. The design of instream structures such as rock and log dams and deflectors used for habitat improvement goes back to at least the 1880s in the United States and even earlier in Europe (Thompson and Stull, 2002). Many of these structures are still used today with very little modification of initial designs, although systematic examination indicates that such structures do not necessarily ensure demonstrable benefits for fish communities (Thompson, 2006), and may in fact decrease habitat abundance and diversity over a period of many decades (Thompson, 2002) (Figure 7.10). In other words, because we typically do not objectively and systematically evaluate the success of restoration projects over periods of several years following project completion, we are not learning from our mistakes.

Small- to medium-scale river restoration has become an industry with designs developed by those

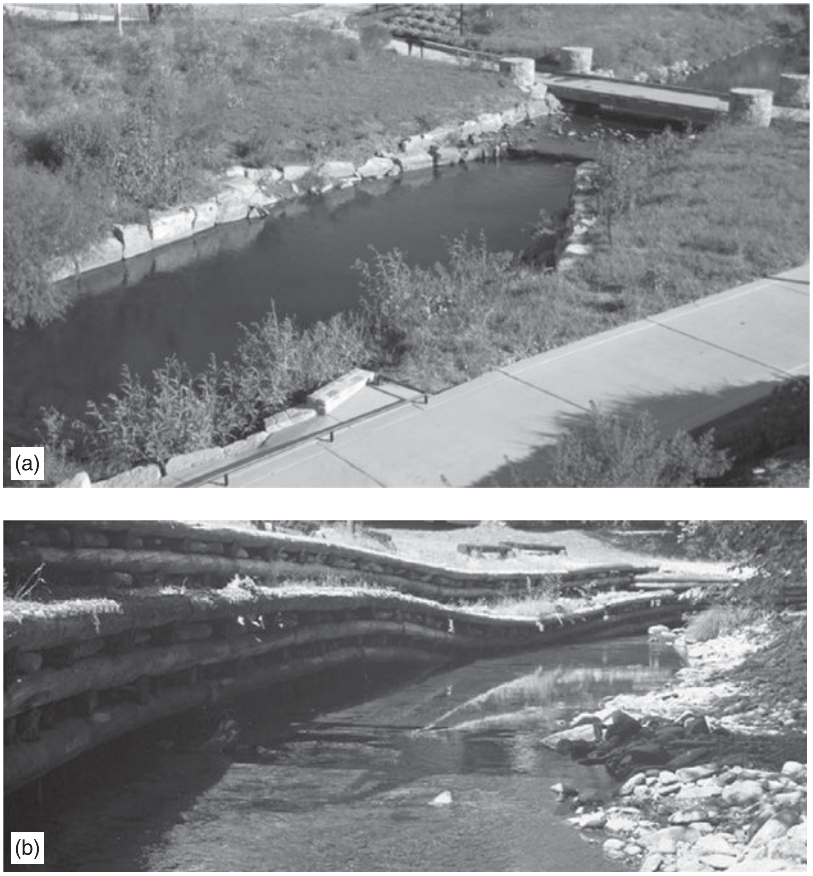

Figure 7.10 Instream structures used for restoration. (a) Vortex weir along a river in Charlotte, North Carolina, USA. A vortex weir is u- or v-shaped with the apex pointing downstream. The structure is designed to deflect toward the channel center and promote bed scour that forms a pool. (b) Collapsed lunker in the Catskills region of New York State, USA. Lunkers are designed to stabilize stream banks and promote edge cover for fish. Both photographs courtesy of Douglas M. Thompson.

with relatively little knowledge of river process and form that are implemented by consulting firms most likely to have a background in civil engineering works rather than in river science. Under these circumstances, the scientific community has become increasingly vocal in criticizing restoration practices (Pasternack, 2013). Numerous papers emphasize that river restoration must be based on or include five factors (Kondolf and Larson, 1995; Hughes et al., 2001; Kondolf et al., 2001; Ward et al., 2001; Hilderbrand et al., 2005; Wohl et al., 2005; Kondolf et al., 2006; Sear et al., 2008; Brierley and Fryirs, 2009; Hester and Gooseff, 2010).

First, restoration should be designed with explicit recognition of complexity and uncertainty regarding river process and form, including the historical context of variations in process and form through time. A well-documented example of failed restoration imposed a stabilized single-thread channel on a river segment that had repeatedly alternated between multi- and single-thread planforms during previous decades in response to fluctuations in flood magnitude and frequency (Kondolf et al., 2001). The “restored” channel was completely altered by a flood within 3 months of project completion.

Second, restoration should emphasize processes that create and sustain river form, rather than imposition of rigid forms that are unlikely to be sustainable under existing water and sediment regimes. Perhaps the most egregious and common example is braided river segments that are "restored" to sinuous, single-thread rivers without addressing the water and sediment yields that produced a braided planform, and without any consideration of meander dynamics (the constructed bends are commonly stabilized to prevent migration).

Third, monitoring of projects after completion must be included, using the set of variables most effective for evaluating achievement of project objectives, and at the correct scales of measurement. If the primary objective of restoration is to enhance biodiversity, for example, then monitoring habitat heterogeneity (under the assumption that habitat heterogeneity always correlates with biodiversity) is not as appropriate as directly monitoring metrics of biodiversity (Palmer et al., 2010b).

Fourth, consideration of the watershed context, rather than an isolated segment of river, is crucial because of the influences of physical, chemical, and biological connectivity on alterations undertaken for river restoration. The Carmel River in California, USA, provides an example where restoration using native riparian vegetation was not initially successful because decades of groundwater withdrawal had lowered the water table below a depth that could be accessed by the vegetation (Kondolf and Curry, 1986). Consequently, the native plants had to be artificially irrigated to ensure their survival.

Fifth, accommodation of the heterogeneity and spatial and temporal variations inherent in rivers is necessary (Brierley and Fryirs, 2009). Rivers continually adjust parameters such as bedform configuration, bed grain-size distribution and channel width/depth ratio in response to fluctuations in water and sediment yield to the channel. These adjustments are commonly not synchronous or of exactly the same magnitude between distinct reaches of the river. Allowing the channel some freedom to adjust to changes imposed during restoration, as well as changes that will inevitably occur after restoration, increases the likelihood that the objectives of restoration will continue to be met over a period of many years.

Numerous papers also examine how numerical simulations can be used to predict restoration outcomes prior to project implementation (Brooks and Brierley, 2004; Singer and Dunne, 2006). However, as Bernhardt et al. (2007) emphasized in their survey of river practitioners, publishing more scientific studies of river restoration will not by itself change the existing situation. River restoration can only improve through direct, collaborative involvement among scientists, managers, and practitioners.

Such collaborations appear to be an obvious next step, but can be very difficult to achieve. Interdisciplinary scientific teams provide significant challenges because of differences in terminology, conceptual models, qualitative versus quantitative knowledge, and temporal and spatial scales of interest between disciplines (Benda et al., 2002). These challenges multiply when the pool of participants is broadened beyond the scientific community. There is no question, however, that river restoration requires an interdisciplinary approach. As reviewed by Pasternack (2013), wetland rehabilitation provides a model. Wetland rehabilitation is facilitated by a certification program hosted by the Society of Wetland Scientists, which includes research scientists, governmental regulators, and practitioners. There is no equivalent for river restoration, partly because the river science community is diverse and oriented toward specific academic disciplines, as well as being strongly divided between research scientists and practitioners, and poorly organized (Castro, 2008; Pasternack, 2013). Although regulators and practitioners would like to establish a universal restoration approach that would standardize methods, research scientists remain highly skeptical that such a "cookbook" technique can be effective. There presently is no scientific consensus about the scientific foundations for restoration, what the practice should entail, or who should be allowed to undertake restoration (Darby and Sear, 2008; Pasternack, 2013).

An important consideration in river restoration is that restoration is not an "all or nothing" process. A river does not have to be—and typically cannot be—restored to some completely natural condition that existed prior to intensive resource use. Partial restoration within the constraints existing in a watershed can nonetheless restore a great deal of physical and ecological form and function. Small watersheds in the central United States that are effectively completely devoted to agriculture, for example, have highly channelized streams. These streams cannot be restored in the traditional sense. River engineering is ubiquitous and occurred prior to scientific study of these streams, so that little is known of reference conditions. In addition, land cover has been altered throughout the watershed, so that water and sediment yields are completely altered (Rhoads et al., 1999). Physical and ecological function of these streams can be improved with *naturalization*, however, which defines a viable management goal for watersheds within landscapes intensively modified by humans. A naturalized channel has sustainable hydraulic and morphologic diversity that supports greater biodiversity than an unrestored channel, even though the naturalized channel may still be completely altered relative to its natural state. A naturalized channel might be deepened but not straightened, for example, allowing the development of limited sinuosity and associated physical diversity in hydraulics, substrate, and channel geometry (Rhoads et al., 1999).

River restoration is also commonly spatially constrained by existing land ownership and use. Scientists undertaking basin-scale restoration in the Missouri–Mississippi River drainage of the United States refer to a *"string of beads" approach* in which land acquisition and restoration activities focus on key floodplain habitats such as flood-prone areas near tributary confluences or remnant backwaters that form beads along the string of the otherwise altered river corridor (Galat et al., 1998). This spatially discontinuous approach to river restoration yields demonstrated improvement in water quality, flood hazard mitigation, and biodiversity.

The European Water Framework Directive provides an example of a new governmental approach that may help to organize efforts toward common objectives and enhance river restoration at a national or transnational scale (European Commission, 2000). The directive is designed primarily to improve and protect water quality, with a set deadline for achieving "good status" for all surface waters within member nations by 2015. This status is to be achieved by meeting requirements for ecological protection and minimum water quality standards analogous to those enforced by the US Environmental Protection Agency. These goals, although laudable, are difficult to implement: 95% of England's rivers, for example, are at risk of failing legislated environmental objectives (Green and Fernández-Bilbao, 2006; Pasternack, 2013).

Supplemental Section 7.3.2 discusses examples of river restoration from diverse river basins, including the Danube.

7.3.3 Instream, channel maintenance, and environmental flows

A vital aspect of river restoration at many sites is preserving or restoring a natural, as opposed to regulated, flow regime. A widely cited paper, "The Natural Flow Regime" (Poff et al., 1997), outlines the importance of magnitude, frequency, duration, timing, and rate of change of flow in river networks for physical process and form and for ecological communities. When flow regime is altered directly by flow regulation or indirectly by changes in land cover that alter water yield to a river, channel characteristics and aquatic biota are affected. Subsequent papers have documented that regulation tends to homogenize river flow regimes, with the consequence that riverine physical characteristics and biotic communities also become more homogeneous with time (Moyle and Mount, 2007; Poff et al., 2007). Growing awareness of the importance of all aspects of a river's flow regime led to the current emphasis on environmental flows.

Initial efforts to protect river flow focused on minimum flows. In arid and semiarid regions such as the western United States, dams and diversions designed to manipulate water for consumptive uses, including agricultural irrigation, can result in river segments that are completely dewatered for some or all of the year. As two fish biologists wrote in one of their papers about rivers in such regions, "it is obvious that without water, there can be no fish" (Fausch and Bestgen, 1997).

The concept *of instream flows* developed as a means to preserve some minimum flow level within the channel. An early version of this was based on the instream flow incremental methodology (IFIM) (Bovee and Milhous, 1978). IFIM uses a biological model that describes the habitat preference of individual fish species in terms of depth, velocity, and substrate, and a hydraulic model that estimates how habitat availability varies with discharge. The intent behind this method is to be able to specify the minimum flows below which individual species likely cannot be sustained in a river segment, as well as the inferred gains in habitat and potentially in fish biomass as flow increases. Although the method has been criticized for being overly simplistic—the biological model does not account for interactions such as competition or predation, for example—IFIM remains widely used in evaluating alternative water management options (Stalnaker et al., 1995).

Minimum flows can allow river organisms to survive for a period of time, but a river that in essence has only continual base flows will eventually lose its capacity to support a diverse aquatic community. Periodic high flows are indirectly necessary to biotic communities because higher flows maintain habitat by performing functions such as scouring pools, winnowing fine sediments from the bed, and limiting channel narrowing through encroachment of riparian vegetation. High flows are also directly necessary because they provide a window of opportunity during which organisms can disperse longitudinally and laterally, accessing new habitat for breeding and feeding.

Recognition of the importance of higher flows first led to the concept of *channel maintenance flows*, typically defined as the components of a river's flow regime necessary to maintain specific physical channel characteristics, such as sediment transport or flood conveyance. Channel maintenance flows are an applied equivalent of the concepts of bankfull, effective, or dominant discharge. In this applied context, channel maintenance flows can specify a particular magnitude and frequency of flow to achieve a limited objective, such as pool scour, or they can incorporate a broader range of flow magnitudes designed to maintain a physically diverse channel (Andrews and Nankervis, 1995). Analogous to instream flows, the intent behind channel maintenance flows is to quantify and then legally establish or protect the magnitude and frequency of flow necessary to maintain specific components of river process and form.

The latest iteration in this progressively expanding view of the components of the flow regime necessary to preserve physical and ecological integrity in a river is environmental flows. The concept of environmental flows grew out of experimental flow releases from dams, such as those on the Colorado River through the Grand Canyon in 1996, 2004, and 2008 (Melis, 2011). An experimental flood is tied to a qualitative or quantitative model of a river ecosystem that predicts some beneficial effect from the flood. In the case of the Colorado River, the floods are designed primarily to deposit finer sediments (silt and sand) along the channel margins in order to restore riparian and backwater habitat that has been lost through progressive erosion of sand bars and backwater habitats since construction of Glen Canyon Dam in 1963. Ideally, the experimental flow release is conducted in an *adaptive management* context in which the results of the flood are systematically assessed, and the underlying model of the river ecosystem is modified as needed (Melis, 2011). Experimental releases from dams are now documented for several rivers in diverse settings (Mürle et al., 2003; Konrad et al., 2011).

Environmental flows now refer both to such experimental releases, which are typically limited in duration, and to an annual hydrograph that specifies magnitude, frequency, timing, duration, and rate of change in flow, commonly for diverse conditions such as wet years and dry years (Rathburn et al., 2009) (Figure S7.8). As with other forms of river restoration, developing the guidelines for environmental flows is time-consuming and challenging because this process forces geomorphologists and riverine ecologists to specify flow thresholds related to targets such as winnowing fine sediment, mobilizing the entire bed, creating overbank flows for channel–floodplain connectivity, or maintaining diversity of species and individual ages within a riparian forest. The complexity and uncertainty associated with river process and form, as discussed at length in Chapters 3, 4 and 5, mean that this task is sometimes straightforward, but more often includes a great deal of uncertainty. Uncertainty that can be acceptable in scientific research becomes more challenging in a management context in which every cubic meter of water released from a hydroelectric dam during an experimental flood, for example, equates to a loss of revenue for the entity operating the dam.

The procedure of assessing environmental flow requirements has gradually assumed characteristic steps of (i) quantifying natural and altered stream flows and the changes in the river flow regime in terms of relevant hydrologic metrics (Richter et al., 1996; Gao et al., 2009) and (ii) quantifying relationships between hydrologic metrics and physical and biological river attributes, which essentially involves coupling physical and biological models (Sanderson et al., 2011). Once recommendations are developed for environmental flows, the process moves into the policy arena in which a community much broader than scientists typically weighs in on water availability and use. As Arthington and Pusey (2003) described the process in an Australian context, the two vital questions are: "How much water does a river need? and How can this water be clawed back from other users?" (p.377). An extensive literature has come into being during the past decade that describes environmental flow assessments and recommendations, as well as a variety of case studies (Tharme, 2003; Arthington et al., 2006; Shafroth et al., 2010).

Environmental flows are in many cases driven by the need to preserve endangered species, and ecologists tend to focus on flow regime. Geomorphologists increasingly emphasize the equal importance of sediment dynamics in maintaining channel complexity, habitat heterogeneity, and nutrient cycling (Pitlick and Wilcox, 2001). Altered flow regimes are commonly accompanied by altered sediment dynamics as a result of sediment trapping behind dams or changed ability of flows to entrain and transport sediment present along the river corridor. A body of scientific literature on sediment dynamics in the context of environmental flows is just beginning to appear (Rubin et al., 1998; Wiele et al., 2007).

7.4 River health

River restoration in any form is driven by the perception that a river is to some extent unhealthy and can be improved. The concept of river health is intuitively appealing to many people and easy to communicate at a general level to non-scientists (Karr, 1999). Scientists debate whether such a conceptualization is useful or appropriate, however, as well as how to quantify river health (Boulton, 1999; Fairweather, 1999; Harris and Silveira, 1999). Much of this debate occurs in the biological literature, partly because river health is an example of ecosystem health (Norris and Thoms, 1999).

River health is related to *ecological integrity*, which is the ability of an ecosystem to support and maintain a community of organisms with species composition, diversity, and functional organization similar to those within natural habitats in the same region (Parrish et al., 2003). This definition of ecological integrity emphasizes biota, but implicitly includes physical and chemical processes that sustain the biota.

Biologists typically explicitly include physical and chemical aspects of rivers in the consideration of river health, as exemplified by defining *river health* as the degree to which a river's energy source, water quality, and flow regime, plus the river's biota and their habitats, match the natural condition at all scales (Karr, 1991; Harris and Silveira, 1999). Numerous qualitative and quantitative metrics of river health have been developed, typically focused on biological metrics (Harris and Silveira, 1999; Karr, 1999), although cumulative metrics sometimes include measures of water quality (Bunn et al., 1999), habitat (Maddock, 1999; Norris and Thoms, 1999), or flow regime (Richter et al., 1996).

Many of the biological metrics used to characterize river health focus on some aspect of biodiversity.

Biodiversity is typically defined in terms of number of species within a given ecosystem, but can be quantified in a wide variety of ways, each of which provides specific information about the ecosystem under consideration. Biodiversity reflects biological influences such as competition and predation, as well as physical influences such as the diversity, abundance and stability of habitat, and the connectivity of habitat (Gaston and Spicer, 2004).

Geomorphologists have been slower to develop metrics of physical river condition to facilitate quantification of difference between contemporary and reference conditions for a river, as well as evaluation of river health. Geomorphic conceptualizations of rivers emphasize diversity of form and process through space and through time (McDonald et al., 2004; Brierley and Fryirs, 2005). Diversity of form and process reflects the hydrology, sediment supply, hydraulics, substrate, geomorphic history, and biota at the site. A desirable level of physical diversity can constitute *physical integrity*, which Graf (2001) defines as a set of active fluvial processes and landforms such that the river maintains dynamic equilibrium, with adjustments not exceeding limits of change defined by societal values. In other words, a river has physical integrity when river process and form are actively connected under the current hydrologic and sediment regime.

A geomorphic perspective on river health would characterize a healthy river as having two basic characteristics. First, the ability to adjust form and process in response to changes in water and sediment yield, whether these changes occur over many decades to centuries (e.g., climate variability) or relatively short time periods (e.g., a large flood or landslide). Second, a healthy river has spatial and temporal ranges of water and sediment inputs and river geometry similar to those present under natural conditions. Discussions of how to evaluate river health have taken on increased importance as broad governmental regulations such as the European Union's Water Framework and Habitats Directive mandate delineation of diverse aspects of river health (Newson and Large, 2006).

One danger involved in this type of national or transnational evaluation is that of over-simplifying what constitutes a natural or healthy river. Some rivers are naturally depauperate in species, for example, because of harsh physical or chemical conditions or a history of geographic isolation. Some rivers receive large sediment inputs and exhibit substantial channel instability because of natural factors such as semiarid climate or erodible lithology in the watershed. Recognition of complexity and diversity as inherent properties within a river network and between river networks remains crucial for both research and management of rivers. This takes us back to viewing rivers in the context of the greater landscape.

7.5 Summary

A diverse array of human activities have indirectly and directly altered the supply of water, sediment, nutrients, and contaminants to rivers, as well as altering channel geometry, fluxes of materials and organisms, and the six degrees of connectivity between a river and the greater landscape. Even regions that have never experienced dense human populations or intensive resource extraction, such as some high-latitude or high-altitude regions, are now affected by atmospheric warming and associated changes in hydrologic balance. An appropriate way to recognize that we live during the Anthropocene is to assume by default that any particular environment has been altered by human activities (Wohl, 2013c).

A major challenge for fluvial geomorphologists is to effectively integrate our understanding of river process and form into contemporary efforts to restore rivers and to assess and protect river health. Among the unique contributions that geomorphologists can make to river restoration are (i) recognition of the historical context of landscapes, including historical human alterations of riverine ecosystems, (ii) knowledge of connectivity and thresholds that influence river response to natural and human-induced alterations, and (iii) ability to quantify thresholds, alternative stable states of a river, landscape resilience, and physical integrity of the critical zone (Wohl, 2013c).

Chapter 8
Rivers in the landscape

This final chapter looks at interactions between rivers and landscapes across varying time and space scales, starting with interactions over geological timescales of 10^3–10^7 years and entire continents. The significance of spatial context and the potential for relatively abrupt spatial transitions in process and form within a drainage basin are examined in the section on geomorphic process domains. The third section of the chapter returns to the idea of connectivity, which was introduced in the first chapter, and explores the implications of connectivity for the diverse river processes and forms discussed throughout this volume. The next section of the chapter explores distinctive climatic signatures of river process and form associated with high latitudes, low latitudes, and warm drylands, respectively. Five regional examples are then used to illustrate how details of climate, geology, and human resource use through time influence river process and form and create a context that must be considered if river management is to be successful.

Preceding chapters have introduced the basic processes of water, solute, and sediment movement into and through channel networks, and the resulting river forms and adjustments through time. This final chapter returns to the idea that river process and form reflect not only physics but also distinctive processes and forms associated with a specific geographic location and the history of that location.

8.1 Rivers and topography

Having examined the details of how water and sediment move down hillslopes and into channels, and then move through a river network, it is useful to step back and consider the larger scale distribution of rivers across continents. Important insights into interactions between rivers and topography can be gained by examining river configuration.

The manner in which rivers both respond to and shape surrounding topography has been investigated systematically for more than a century. Early work explored why some rivers cut through mountain ranges rather than simply flowing downward from high points in the landscape. Significant questions at the time of this late-nineteenth-century research included, What is the role of rivers (as opposed to glaciers or other processes) in cutting deep river canyons?, and How do large-scale structures (mountains, deep canyons) relate to movements of Earth's crust? Subsequent research has emphasized (i) how redistribution of mass at the surface by river erosion can influence redistribution of subsurface mass via movements of molten material in the crust and tectonic movements and (ii) how

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river gradient and channel width can be used as indicators of spatial variations in rock uplift.

The most obvious topographic influences on river process and form occur in mountainous environments where rivers cut deep, narrow canyons as they flow down toward adjacent lowlands. For more than a century, however, investigators have recognized that rivers do not always follow topography. Observing rivers that cut across mountain ranges in the western United States, Powell (1875, 1876) distinguished *antecedent* drainage networks in which pre-existing channels maintained their spatial arrangement while the underlying landmass was deformed and uplifted, and *superimposed* channels which incised downward to a buried structure. In either case, the river flowed through or across the mountain range, rather than being a consequence of the topography. Today these two types of drainages are commonly referred to as *transverse drainages* (Figure S8.1) that cut across bedrock topographic highs such as anticlines or upwarps (Douglass and Schmeeckle, 2007). Transverse drainage subsumes antecedent and superimposed drainages, overflow, and drainage piracy. In *drainage piracy*, a drainage network on one side of a bedrock high erodes headward sufficiently to lower or breach the drainage divide and divert flow from a network on the other side of the divide. Continuing research has developed tools that illuminate the influences of tectonics on the spatial arrangement of rivers and the geometry of individual river channels, as well as the influences of river incision on tectonics and topography.

The gradient and width of rivers incised into bedrock are the geometric parameters most commonly used to infer the spatial distribution and relative magnitude of tectonic forces. A river incised into bedrock, rather than alluvium, implies that the channel's capacity to transport sediment exceeds the sediment supply (Howard, 1980). Regardless of where they occur along a river's course, bedrock river segments typically have smaller width/depth ratios and steeper gradients than alluvial segments (Montgomery and Gran, 2001; Wohl and David, 2008). These differences in geometry effectively enhance the flow's limited ability to incise the channel bed and enlarge the channel cross section relative to alluvial segments.

Bedrock river segments can be interpreted as geologically transient features that are not confined to headwaters. The middle and lower Danube River of Europe alternately flows across large alluvial basins and then cuts through mountain ranges. The lower Mississippi River, commonly thought of as a fully alluvial river, is better described as a mixed bedrock-alluvial channel because of the presence of a cohesive, Pleistocene-age clay unit that influences river process and form in a manner analogous to bedrock (Schumm et al., 1994; Nittrouer et al., 2011).

The relative lack of erosive ability that produces a bedrock river segment can reflect greater erosional resistance where the river crosses a different lithology (Wohl, 2000b), or changes in relative base level associated with base level fall or with uplift of the drainage (Howard, 1980; Seidl et al., 1994). Where the presence of bedrock river segments reflects enhanced incision in response to relative base level fall, river incision is the primary non-glacial mechanism of transmitting base level changes across the landscape (Hancock et al., 1998). Bedrock river incision steepens adjacent hillslopes, increases topographic relief between summits and valley bottoms, removes mass from the landscape, and ultimately sets the rate at which the entire landscape evolves (Howard, 1994; Burbank et al., 1996; Hancock et al., 1998) (Figure 8.1).

The idea that hillslopes and rivers mutually adjust was first expressed by G.K. Gilbert (1877), and subsequently formally stated in the context of *dynamic equilibrium* by Hack (1960) as a condition in which "… every slope and every channel in an erosional system is adjusted to every other. When the topography is in equilibrium and erosional energy remains the same all elements of the topography are downwasting at the same rate" (Hack, 1960, p. 80).

8.1.1 Tectonic influences on river geometry

Increased availability of topographic data in the form of electronic digital elevation models (DEMs)

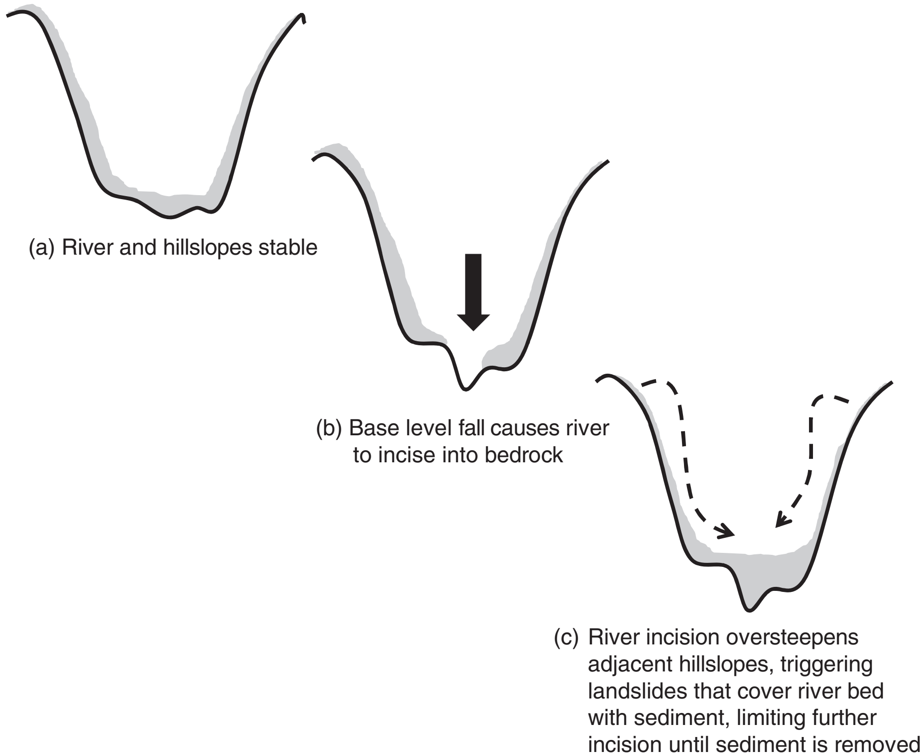

Figure 8.1 Illustration of the effects of bedrock channel incision on landscapes. In these schematic views looking upstream within a river valley, base level change triggers river incision that affects the stability of adjacent hillslopes and tributaries. Gray shading represents hillslope regolith and valley alluvium.

greatly enhanced the ability to detect irregularities in river longitudinal profile starting in the 1990s. Profile irregularities (Figure S8.2) can reflect downstream variations in lithology and erodibility (Valla et al., 2010), glacial history (Hobley et al., 2010), sediment inputs (Cowie et al., 2008), and rock uplift (Snyder et al., 2000), so interpreting the significance of irregularities typically requires knowledge of other characteristics of the river environment. Where investigators have independent evidence of uplift rate, as in the central Apennines of Italy (Whittaker et al., 2008) or the Santa Ynez Mountains of California (Duvall et al., 2004), steeper gradients strongly correlate with greater rates of rock uplift (Whipple et al., 2013).

Spatial variations in channel width/depth ratio are not as readily detected using remote information as are changes in river gradient, but variations in the width of bedrock channels can also reflect differential uplift (Whittaker et al., 2007a, 2007b; Attal et al., 2008; Yanites et al., 2010) as well as changes in rock erodibility (Wohl and Merritt, 2001). Typically, segments of higher uplift or more resistant rock have deeper, narrow cross-sectional geometry.

Adjustments of gradient and width in response to increasing substrate resistance or uplift are typically tightly coupled (Whipple, 2004; Stark, 2006). The most commonly used approach is to interpret downstream variations in scaling relations among channel width w, drainage area A, gradient S, and discharge Q—in other words, downstream hydraulic geometry—as reflecting changes in rock erodibility or uplift rate (Duvall et al., 2004; Cowie et al., 2006; Jansen, 2006). Scaling laws change along bedrock channels crossing an active fault in the central Italian Apennines, for example (Whittaker et al., 2007a), and along rivers crossing growing folds in New Zealand (Amos and Burbank, 2007).

8.1.2 Effects of river incision on tectonics

Decades of research indicate that rivers can convey tremendous volumes of sediment from high-relief landscapes. Rivers remove up to five times more sediment per unit area from mountainous basins than from lowland basins (Corbel, 1959; Willenbring et al., 2013). Milliman and Syvitski (1992) estimated that 96% of the approximately 7819 million tons of sediment delivered to the oceans by rivers each year originates in mountainous settings.

By the 1990s, investigators realized that one implication of this ability to remove mass from mountainous regions is that river incision can affect crustal structure in mountain belts by changing the distribution of stress in the crust (Molnar and England, 1990; Hoffman and Grotzinger, 1993; Beaumont and Quinlan, 1994; Small and Anderson, 1998). Local rheological variations arise in a deforming orogen as a result of deep and rapid incision by glaciers or rivers (Zeitler et al., 2001). The crust weakens as the strong upper crust is locally stripped from above by erosion. This causes the local geotherm (or rate of change in temperature with depth below the surface) to steepen from below as a focused, rapid uplift of hot rock occurs. In other words, incision by glaciers or rivers removes enough mass that molten material rises preferentially into the eroding area from Earth's interior. If efficient erosion continues, material continues to flow into the weakened zone, maintaining local elevation and relief (Koons et al., 2002; Booth et al., 2009a, 2009b). This conceptualization of the interactions between river erosion, uplift, and topography is known as the *tectonic aneurysm model* (Zeitler et al., 2001).

A river's ability to incise depends partly on discharge and the climate that supplies runoff, and the picture now emerging emphasizes strong coupling among climate, erosion, and tectonics. Gradients in climate (Bookhagen and Burbank, 2010) and tectonic forcing influence erosional intensity, which governs the development of topography, which in turn influences climate and tectonics (Roe et al., 2002). This is demonstrated in the Himalayas (Montgomery and Stolar, 2006), where erosion along major rivers causes focused rock uplift. The uplift creates anticlines and the anticlines correlate with local rainfall maxima because monsoon precipitation is advected up the river valleys. The greater rainfall in turn increases the erosive capability of the rivers (Figure 8.2).

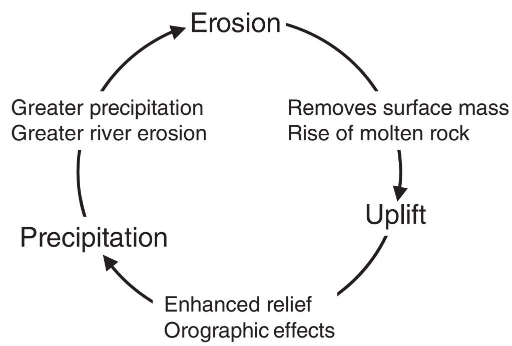

Figure 8.2 Schematic illustration of the interactions among tectonics, topography, and climate, as illustrated by research in the Himalaya.

8.1.3 Indicators of relations between rivers and landscape evolution

Davis (1899) first attempted to relate river geometry to landscape evolution in a conceptual model known as the *cycle of erosion*. This model posited high topographic relief and rivers with steep gradients in geologically young landscapes. As erosion gradually transferred mass to lower elevations, topographic relief and river gradients progressively decreased from mature to old landscapes.

The cycle of erosion, despite the name, assumed a highly linear landscape evolution with time and implied that an observer could readily interpret the relative geologic age of distinct landscapes based on their topography. This conceptualization is typically contrasted with the ideas of Davis' contemporary, G.K. Gilbert. Gilbert emphasized nonlinear change with periods of little net change, or equilibrium, as a result of feedbacks such as those subsequently recognized as tectonic aneurysms or isostatic rebound (Gilbert, 1877). *Isostatic rebound* is delayed upward flexure of Earth's crust in response to removal of mass such as a continental ice sheet that had depressed the crust when the ice sheet was present. The implication of tectonic aneurysms or isostatic rebound is that elevation or relief may change relatively little over time spans of 10^3–10^4 years, despite continued erosion and transfer of mass to lower elevations.

Many subsequent investigators have demonstrated that rates of landscape change fluctuate substantially through time and space (Bierman and Nichols, 2004). Conceptual models now tend to emphasize that most landscape changes occur during relatively short periods of time and are concentrated in relatively small portions of a drainage basin, although net change in elevation or relief may be minor.

Topographic metrics are still used to infer relative rates or stages of landscape evolution. Among these metrics are *hypsometric curves* (Strahler, 1952), which illustrate the distribution of mass within a basin by plotting proportion of total basin height against proportion of total basin area (Figure 8.3). Strahler (1952) proposed that these curves could be used to distinguish relative basin age as a function of decreasing hypsometric integral—the area under the hypsometric curve—with increasing age. Subsequent research suggests that hypsometric curves can be used to infer the history and processes of basin development. The distribution of mass within a basin reflects uplift rates and variations in erodibility of different lithologic units (Walcott and Summerfield, 2008; Pérez-Peña et al., 2009), as well as differences in diffusive (hillslope) versus fluvial sediment transport (Willgoose and Hancock, 1998) and glacial versus fluvial erosion (Sternai et al., 2011). Hypsometric curves are likely to be concave-down everywhere, for example, within landscapes dominated by diffusive transport (Willgoose and Hancock, 1998), and glacial valleys are more likely to have concave-up curves than are fluvial valleys (Sternai et al., 2011).

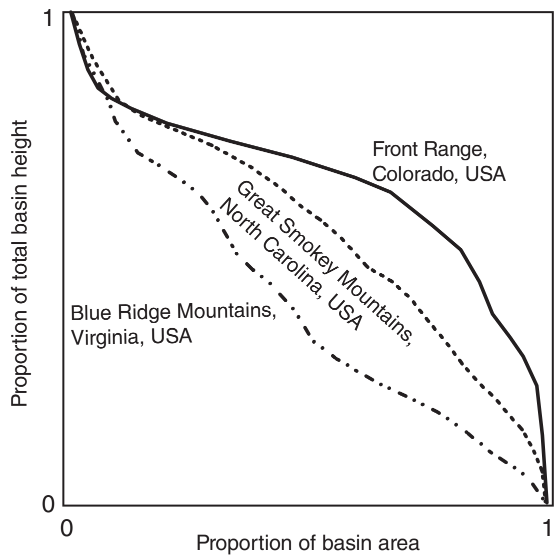

Figure 8.3 Hypsometric curves for three small drainage basins in mountainous regions of the Colorado Front Range, the Blue Ridge Mountains of Virginia, and the Great Smokey Mountains of North Carolina, all in the United States. The river in the Colorado Front Range begins on a broad, low-relief upland and then descends steeply through a narrow canyon. The catchment in the Great Smokey Mountains is well dissected, with no alluvial fans, no floodplains, and minimal deposition in the valleys. The catchment in the Blue Ridge Mountains has alluvial fans at the base of the ridges and extensive valley infill, so that more of the mass is at lower elevations in this catchment. Figure courtesy of Gregory S. Springer. (From Wohl, 2010, Figure 2.30.)

8.1.4 Tectonics, topography, and large rivers

Although much of the research summarized in the preceding sections has focused on rivers in mountainous terrain, investigators have also examined the spatial arrangement of very large rivers in the context of tectonic history and topography. Potter (1978) described how the location and configuration of most of Earth's largest river basins reflect the tectonic assembly and deformation of continental land masses. He noted that the basic configuration of some major rivers has been extremely persistent. The Mississippi drainage, for example, has existed about 250 million years, or about 1/16 of Earth's history (Potter, 1978). This persistence partly reflects the control of deeper crustal structures.

The 28 largest rivers discharge across trailing-edge coasts without compressional deformation, which reflects the continental asymmetry of many watersheds (Inman and Nordstrom, 1971). The mouths of many large rivers occupy grabens or crustal downwarps. The alignment of several major rivers, including some in South America (Potter, 1978), the Nile (Schumm and Galay, 1994), and the Rio Grande in North America, correlate with large-scale crustal fracture patterns.

The Amazon River provides an example of the three scales at which tectonics can influence large-scale features of a drainage network (Mertes and Dunne, 2007; Dunne and Aalto, 2013). At the continental scale (5×10^3 km), the assembly of orogen (mountains), foreland basin, cratons, and grabens influences production of runoff, sediment supply, and accommodation space. Specifically, the Amazon heads on the Andean arc at the leading edge of the South American Plate. The Andes provide the majority of the sediment that the Amazon carries across a very broad lowland before entering a graben that localizes the mouth of the river on the trailing edge of the continent.

At an intermediate scale (10^2–10^3 km), the spacing of crustal warping transverse to the river course influences gradient, valley width, channel sinuosity, accommodation space, and sediment distribution across the floodplain. Four major structural arches lie transverse to the main course of the Amazon between the Peru–Brazil border and the Atlantic Ocean (Mertes and Dunne, 2007). As the Amazon crosses each of these arches, channel gradient increases, the valley grows narrower, and the channel becomes less sinuous.

At the local scale (10^1–10^2 km), brittle crustal fracturing influences channel orientation and gradient. Channel alignment follows fractures because these create localized zones of more readily eroded bedrock (Latrubesse and Franzinelli, 2002).

These examples illustrate that, when examining river process and form, it is important to remember that deeper crustal structures and geologic processes occurring over millions of years can strongly influence river characteristics. These influences may be most readily detected in mountainous portions of a river network, but can also influence the world's largest lowland rivers, such as the Amazon or the Mississippi.

8.2 Geomorphic process domains

Turning to smaller spatial scales at the subwatershed or reach level, distinct suites of geologic and climatic processes can create spatial differentiation within river networks. Schumm (1977) conceptualized drainage basins as consisting of an upstream zone of production from which water and sediment are derived, a central zone of transfer in which inputs can equal outputs in a stable river, and a downstream zone of deposition (Figure S2.20). Although acknowledging that production, transfer, and deposition occur continuously throughout a drainage basin, this organization recognizes the existence of spatial zonation in dominant processes within a catchment.

Subsequent conceptual frameworks have also emphasized spatial zonation. Montgomery and Buffington (1997), for example, distinguished source, transport, and response segments in their reach-scale classification of mountain channel morphology (Figure 4.9). Sklar and Dietrich (1998) hypothesized consistent changes in dominant incision mechanism (debris flow, fluvial) and substrate type (coarse bed alluvial, fine bed alluvial) at threshold slopes, regardless of drainage area (Figure 2.11).

Montgomery (1999) built on this work in describing *process domains*, which he defined as spatially identifiable areas of a landscape or drainage basin characterized by distinct suites of geomorphic processes (Figure 2.8). The existence of process domains implies that river networks can be divided into discrete regions in which ecological community structure and dynamics respond to distinctly different physical disturbance regimes (Montgomery, 1999).

As noted earlier, some river geomorphic parameters exhibit progressive downstream trends, whereas others exhibit so much local variation that any systematic longitudinal trends that might be present are obscured (Wohl, 2010). Local variation that overwhelms progressive trends is particularly characteristic of mountainous terrain, where spatially abrupt longitudinal zonation in substrate resistance, gradient, valley geometry, and sediment sources can create substantial variability in river process and form. Under these conditions, characterizing river dynamics at reach scales can be more accurate than assuming that parameters will change progressively downstream. Examples of geomorphic parameters for which spatial variation is better explained by process domain classifications than by drainage area or discharge in mountainous drainage basins include riparian zone width (Polvi et al., 2011), floodplain volume and carbon storage (Wohl et al., 2012c), and instream wood load (Wohl and Cadol, 2011) (Figure S8.3).

Process domains can also apply to very large drainage basins that have distinct spatial differences associated with topography or climate. The wet, high-relief, sediment-producing Ethiopian Highlands portion of the Blue Nile, for example, is distinctly different than the dry, low-relief mainstem Nile lower in the drainage. Similarly, the steep, narrowly confined segments of the Danube that cut through mountainous terrain are distinctly different than the intervening anastomosing or braided segments in broad alluvial basins.

Process domains can provide a useful organizational framework for delineating the spatial distribution and relative abundance of different valley and channel types (Wohl et al., 2007), and this can facilitate identification of sensitive or rare areas and formulation of different management strategies for distinct physical settings. The concept of process domains can be readily applied at the reach scale at which most river management occurs. With even minimal field calibration, process domains can also provide a framework for remotely predicting at least relative variations in numerous valley-bottom characteristics.

Other conceptual frameworks that emphasize spatial zonation include the *River Styles framework* (Brierley and Fryirs, 2005) structured around the five spatial scales of the

- catchment,
- landscape units of relatively homogeneous topography within the catchment,
- reaches with consistent channel planform, assemblage of channel and floodplain geomorphic units, and bed-material texture,
- channel and floodplain geomorphic units such as pools, and
- hydraulic units of homogeneous flow and substrate characteristics.

River style is classified at the reach scale. The catchment-wide distribution of river styles can be used to understand spatial distribution of controls on river process and form, relative abundance of different river styles, and potential sensitivity and resilience of different portions of the river network, analogous to the application of process domains.

Stream biologists also emphasize zonation at differing spatial scales. This is exemplified by the widely used hierarchy of stream system (10^3 m length), stream segment (10^2 m), reach (10^1 m), pool/riffle (10^0 m), and microhabitat (10^{-1} m) of Frissell et al. (1986).

Aquatic and riparian ecologists have also examined the relative importance of progressive trends versus local controls. One of the primary conceptual models of aquatic ecology, the *river continuum concept*, emphasizes continuous longitudinal gradients in the structure and function of ecological communities along a river network (Vannote et al., 1980). In contrast, the *serial discontinuity concept* focuses on how dams disrupt longitudinal gradients along river courses (Ward and Stanford, 1983, 1995). *Hierarchical patch dynamics* (Pringle et al., 1988; Poole, 2002) emphasizes the existence of relatively homogeneous units from the scale of microhabitat up to channel reaches, with distinct changes in process and form between patches. Reach-scale patches, like process domains or river styles, can result from changes in physical processes such as glaciation or differential rock erodibility (Ehlen and Wohl, 2002) and from biotic influences such as beaver dams (Burchsted et al., 2010).

The appropriateness of conceptual models that emphasize progressive downstream changes versus patchiness varies depending on the river characteristic being described and the specific drainage basin. Recognition that a variety of river processes and forms can exhibit abrupt spatial transitions, however, illustrates the importance of considering landscape context when examining river process and form. The presence and characteristics of connectivity exert an important influence on both spatially continuous and discontinuous processes and forms.

8.3 Connectivity

Although the first chapter of this volume introduced the idea of diverse forms of connectivity, it is worth returning to this concept and exploring the implications of connectivity for several of the processes and forms discussed to this point, including

- interactions between hillslope, floodplain, channel, and hyporheic environments;
- sources, transport, and residence time of water, sediment, and solutes; and
- spatial zonation within a drainage basin.

Figure 8.4 illustrates how various forms of connectivity change throughout a drainage basin.

High connectivity implies that matter and organisms move rapidly and easily within a river network. Landscapes typically include some characteristics that create at least temporary storage and limit connectivity. Subsurface units of low permeability can limit the downslope transmission of water from hillslopes to channels, or limit hyporheic and groundwater exchanges along channels. Lakes, broad floodplains with extensive wetlands, and numerous channel-spanning obstructions such as beaver dams or logjams can substantially decrease the rate at which floods move through a river network. Similarly, depositional features such as alluvial fans can limit the rate at which sediment is introduced from hillslopes to channels. Extensive floodplains can increase the time necessary for sediment entering a river network to move completely through the network. As described in Section 6.1.2, sediment can reside on the floodplains of the Amazon River for thousands to millions of years. Erosionally resistant portions of a river network can influence landscape connectivity. Segments of bedrock river commonly act as local base levels, for example, and the upstream transmission of base level change is limited to the rate at which the river can incise the bedrock segment. Large waterfalls or portions of an ephemeral or intermittent river network that are dry can limit migration of organisms and thus limit biological connectivity (Jaeger and Olden, 2012).

Features that limit connectivity can be conceptualized as reservoirs that store materials, as exemplified by alluvial fans storing sediment or floodplains storing peak flows during a flood. Features that limit connectivity can also be conceptualized as barriers, as in the case of a local base level that limits profile adjustment or a dry stream segment that limits fish dispersal. Whether a reservoir or a barrier, these aspects of river networks exert critical controls on fluxes of material and organisms and must be included when understanding or quantifying all aspects of river networks, from production of water, solutes, and sediment, to movement of these materials downslope into channels and through channel networks.

Among the challenges in managing rivers are those of quantifying connectivity and understanding how human activities have increased or decreased connectivity within a landscape (Kondolf et al., 2006). As discussed in the introductory chapter, most human activities decrease hydrological, sediment, biological, and landscape connectivity within a river network, although a few alterations such as flow diversions and removal of naturally occurring obstructions such as beaver dams can increase connectivity.

Unanticipated side effects of altered connectivity can require many decades to become apparent. Before the 1970 completion of the Aswan High Dam

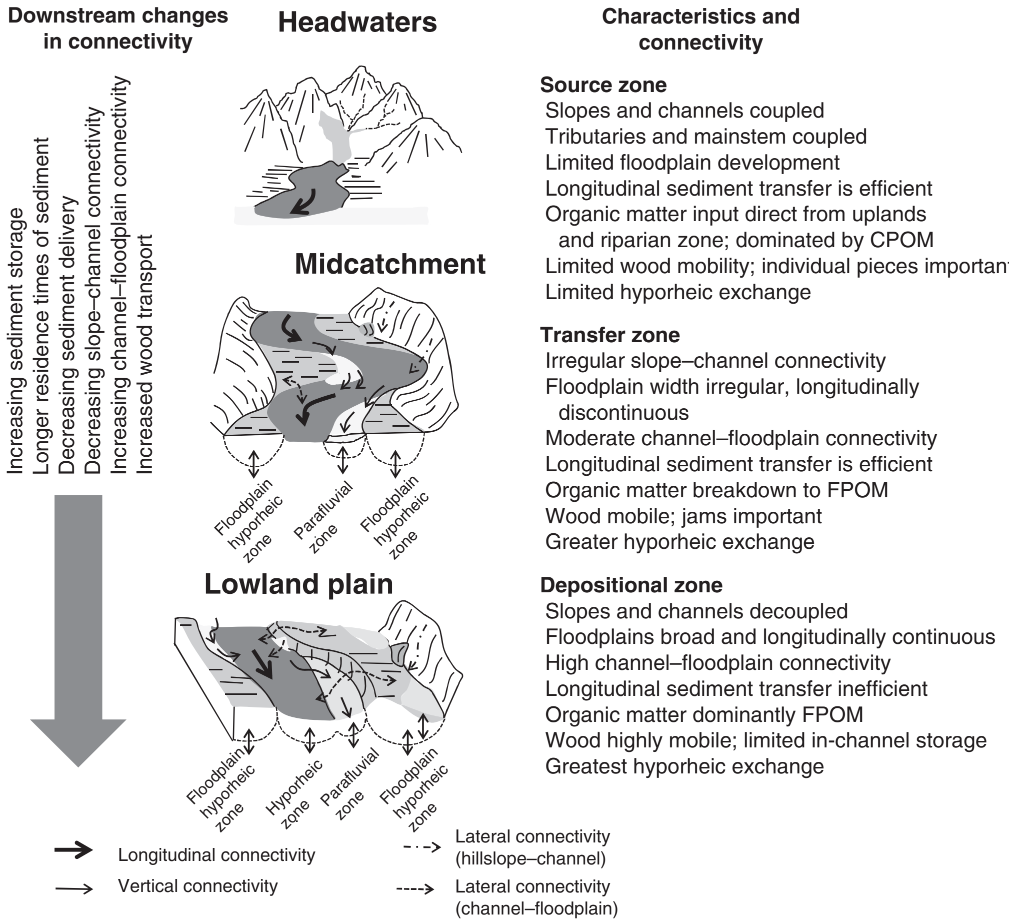

Figure 8.4 A schematic illustration of changes in connectivity with distance downstream along a river with high-relief headwaters. Moving downstream, the river flows through headwater valleys with relatively thin, narrow alluvial veneers over bedrock, and then through progressively wider and deeper alluvial valleys with greater floodplain development and hyporheic exchange. The presence of a floodplain buffers the mainstem river from hillslope and tributary inputs by creating depositional zones along the river, and progressively more extensive floodplains typically equate to greater average residence time of sediment, surface flow during overbank floods, and subsurface flow. CPOM is coarse particulate organic matter (>1 mm in diameter) and FPOM is fine particulate organic matter (0.45 μm–1 mm). (From Brierley and Fryirs, 2005, Fig 2.10, p. 44.)

on the Nile River in Egypt, the river annually carried more than a hundred million tons of silt to the Nile Delta. The dam now traps much of this sediment, causing subsidence and erosion in the delta: former delta villages are now 2 km out to sea. Plankton formerly nourished by nutrients in the river flow have dramatically decreased in abundance, contributing to a collapse of sardine populations that fed on the plankton. Reduction of sediment connectivity along this major river has thus altered the physical, chemical, and biological characteristics of the lower river and the nearshore zone (Wohl, 2011a).

Along Siberia's Ob River, the Novosibirsk Dam reduces seasonal peak flows along more than a thousand kilometers of river. The Ob historically provided an important commercial fishery, but many species of fish require access to the floodplains for spawning and nursery habitat during the spring peak flows. Fish in the Ob River need at least 20 days of flooding in order to spawn, hatch, and grow.

The Novosibirsk Dam has reduced floodplain habitat by half during years of average flow, and eliminated this habitat during dry years. This dam, along with several others in the Ob catchment, also limits longitudinal movements by fish, creating genetic isolation and constraining the ability of each fish population to find appropriate habitat during dry periods. The commercial fishery along the Ob has largely collapsed since construction of the dams (Wohl, 2011a).

The effects of altered connectivity along the Nile and the Ob might appear obvious in retrospect, but the challenge of anticipating not only the type, but the magnitude, of altered connectivity in connection with river engineering remains substantial. Numerical simulation can be particularly useful in this context by facilitating the ability to evaluate alternative scenarios, but any model used must be accurately parameterized and underpinned by a solid understanding of the sources and characteristics of connectivity within a drainage basin. Again, knowledge of landscape context is critical to effective understanding and management of river process and form.

8.4 Climatic signatures

Many of the preceding chapters have briefly mentioned differences in river process and form in relation to climate. Climatically induced diversity in process and form has been relatively neglected in scientific river studies, however, because the great majority of studies have been conducted in the temperate latitudes where many scientists live. This imbalance and relative neglect of high latitudes, low latitudes, and drylands has been changing in recent decades, and this portion of the chapter reviews the implications of climatic differences for interactions between process and form.

8.4.1 High latitudes

The salient feature of high latitudes with respect to interactions between river process and form is the occurrence of very cold temperatures that maintain seasonal ice cover on rivers (Figure S8.4). As discussed in Section 3.2.6, the formation, presence, and break-up of ice create distinct hydrologic and hydraulic effects on stage and discharge. The presence of ice can also strongly influence channel and floodplain erosion, deposition, and connectivity.

Ice cover and ice jams that form during break-up create backwater effects that enhance overbank flooding. When ice jams break, the resulting surge can enhance overbank flooding downstream. Changing hydraulic forces during ice jams and break-up, along with the mechanical effects of large chunks of moving ice, can enhance bank erosion and the overbank flooding and scouring that structure riparian vegetation (Beltaos, 2002). Avulsion induced by ice jams has been described for both meandering and braided channels, and channel ice can facilitate bank erosion by gouging the banks, increasing the bank loading, and reducing vegetation growth along the banks (Ettema and Kempema, 2012). Ice cover can alter lateral variations in flow depth and boundary shear stress within a channel (Ettema and Kempema, 2012). Ice–backwater effects can alter the direction of flow within a channel and the connectivity between the main channel, secondary channels, and the floodplain (Prowse and Beltaos, 2002). Ice that retards flow can decrease bed-material transport, but ice can also increase sediment transport by directly moving the bed sediment via ice rafting (transport of sediment by floating ice), ice gouging, or ice push (Ettema and Kempema, 2012).

Ice jams and the surges that result from their release can cause habitat degradation or loss, species stress or mortality, and deposition of fines and deterioration of spawning grounds. The highest suspended sediment concentrations occur during freeze-over and break-up, however, and ice jams and surges also replenish adjacent floodplains with sediment and nutrients (Beltaos, 2002).

Although there is no unique type of cold-region river geometry, *sandur* (a valley segment undergoing rapid aggradation, with a downstream decrease in particle size), braided channels, meandering channels, and anastomosing channels in wetland environments are particularly common in cold regions (Vandenberghe and Woo, 2002). For all of these channel types, the periods of dynamic change in seasonal ice cover—freeze-over and break-up—tend to be periods of the greatest channel and floodplain geomorphic change in cold-region rivers (Prowse and Beltaos, 2002), even though these periods do not coincide with the greatest seasonal discharge. The enhanced erosion associated with ice also increases the frequency of changes in channel cross-sectional geometry, planform, and channel–floodplain connectivity (Ettema and Kempema, 2012).

8.4.2 Low latitudes

The salient feature of low latitudes is the magnitude and speed with which various fluxes occur (Wohl et al., 2012a). Low latitudes here are synonymous with the tropics, the area of surplus radiative energy bounded by anticyclonic circulations near 30° N and 30° S (Scatena and Gupta, 2013). This region includes the humid tropics, where average annual rainfall is greater than potential evapotranspiration and precipitation is sufficient to support evergreen or semi-deciduous forests, and the dry tropics. The dry tropics are sufficiently similar to temperate dry regions to be treated in the next section.

The humid tropics can be further distinguished as seasonal and aseasonal (Gupta, 1995). The seasonal humid tropics have a marked seasonal concentration of rainfall and runoff, typically reflecting the intertropical convergence zone (ITCZ) or monsoonal circulation patterns, and 80% of the annual runoff can occur in 4 or 5 months of the year (Scatena and Gupta, 2013). This can result in substantial variations in river process and form between wet and dry seasons, including distinctly different wet- and dry-season channel geometry (Wohl, 1992; Gupta, 1995). The aseasonal humid tropics typically have mean annual rainfall between 2000 and 4000 mm/year, whereas mean annual rainfall in the seasonal humid tropics varies between 1000 and 6000 mm/year and interannual variability in runoff can be large (Scatena and Gupta, 2013).

Extremely intense rainfall and preferential shallow flow paths such as macropores (Section 2.2.1) result in large hydrologic inputs to tropical channels, which tend to have a very flashy flow regime in smaller drainages and a prolonged peak flood (exceeding 3 months) in very large basins such as the Amazon and Congo. Smaller catchments are more likely to have basin-wide intense storms than are equivalently sized catchments at higher latitudes. Monsoons, hurricanes, and ITCZ-related storms tend to cover sufficiently large areas that even larger catchments receive geomorphically significant rainfalls at the same time (Scatena and Gupta, 2013). Widespread intense rainfall results in high percentages of contributing area and channel-modifying discharges that occur simultaneously throughout the catchment. These characteristics contrast with the more spatially restricted precipitation inputs and channel modifications of equivalently sized catchments at higher latitudes.

Continual high air temperatures and abundant vegetation combine with high values of precipitation to create rapid weathering of rock and soil and of organic inputs such as wood. Large inputs of material to channels occur in high-relief tropical environments when cyclones or hurricanes trigger widespread landslides that strip hillslopes of weathered rock and vegetation (Scatena and Lugo, 1995; Goldsmith et al., 2008; Hilton et al., 2011b; Wohl et al., 2012b) (Figure S8.5). Instream wood does not persist in low-latitude rivers. Although individual pieces of wood and large jams can create important geomorphic and ecological effects (Wohl et al., 2009, 2012b), the transience of wood as a result of combined rapid decay and high transport rates is particularly noticeable (Spencer et al., 1990; Cadol and Wohl, 2011; Soldner et al., 2004).

Although the hydrology of low-latitude rivers has distinctive characteristics and process–form interactions occur more rapidly and frequently than in higher latitudes, a recent review concluded that tropical rivers do not have diagnostic landforms that can be solely attributed to their low-latitude location (Scatena and Gupta, 2013). A key distinction relative to rivers in higher latitudes is the high frequency of geomorphically significant flows and channel changes.

8.4.3 Warm drylands

The salient feature of interactions between process and form within rivers in warm dryland regions is spatial and temporal discontinuities. As with tropical rivers, a recent review of dryland channels concluded that there are no features unique to this subset of rivers, although certain characteristics are more common in drylands than elsewhere (Tooth, 2013). Warm dryland here includes hyperarid, arid, semiarid, and dry subhumid environments, but excludes cold, high-latitude or high-altitude regions. Warm dryland regions are particularly prevalent within the subtropical, high-pressure belts of the northern and the southern hemisphere, and much of the scientific research on rivers in these regions has been conducted in the western United States, Australia, the Middle East, and southern Africa.

Warm drylands share high, but variable, degrees of aridity that reflect low precipitation and high potential evapotranspiration. Long periods with little or no precipitation and stream flow are interrupted by intense rainfall and runoff that can generate short-duration flash floods. Irregular precipitation and low water tables keep many dryland channels ephemeral or intermittent, although rivers that originate in wetter mountainous highlands before flowing into dry regions can be perennial (Tooth, 2013). Short periods of high water and sediment connectivity between uplands and channels can result from overland flow and limited upland vegetation. Limited riparian vegetation and sediment cohesion can create unstable banks that promote a braided planform, and floodplains are limited in extent.

Flash floods within channels or across piedmonts are particularly characteristic of catchments less than 100 km^2 in size (Tooth, 2013). Flash floods have steep rising and recessional limbs and may last only minutes to hours, but can generate very high values of discharge per unit drainage area and correspondingly substantial erosion and deposition. Larger river networks are likely to have more sustained flows. Downstream transmission losses in dryland channels (Section 3.2), along with limited contributing area where small convective storms affect only a limited portion of a large drainage basin, create large spatial variability in discharge and temporal variability between storms (Tooth, 2013).

Although ephemeral dryland channels can have less dense riparian vegetation than perennial rivers, dryland vegetation can be concentrated in and along channels and can generate substantial flow resistance during flows. Resistance can increase with stage as woody vegetation becomes immersed, particularly along the tops of banks and bars (Knighton and Nanson, 2002). Resistance associated with riparian vegetation can lead to a scenario of streambed scour at lower stages during a flood, when vegetation is not inundated and thus contributes little resistance, and fill during the higher stages when bank flow resistance is strongly affected by vegetation (Merritt and Wohl, 2003).

Dryland rivers typically transport large quantities of suspended and bed load and display strong hysteresis of bed scour during the rising limb and fill during the falling limb. Scour is facilitated by the absence or poor development of coarse surface layers. Poorly developed coarse surface layers may reflect abundant upland sediment supplies, enhanced sediment mixing during scour and fill, high rates of bedload transport, and short-duration flows that minimize winnowing of fine particles from the bed (Tooth, 2013). Lack of coarse surface layers also facilitates high rates of bed load transport that increase more consistently with increasing flow because particles across a large size range are available for entrainment at the start of flow and the bed becomes highly mobile at even modest flows (Tooth, 2013).

Much of the research on small- to medium-size dryland channels emphasizes abrupt transitions between incising and aggrading conditions across a channel network and through time. This emphasis is exemplified by the extensive literature on arroyos in the American Southwest (Graf, 1983, 1988; Harvey, 2008).

Repeated large floods appear to dominate process and form in many dryland channels, as evidenced by numerous studies of flood-related channel changes and recovery times of decades or longer (Tooth, 2013). Dryland channels may be particularly susceptible to change during floods because of limited bank cohesion in the absence of dense riparian vegetation and abundant silt and clay, and because of the lack of intervening smaller flows that could modify flood-generated erosional and depositional features (Tooth, 2013).

In contrast to the emphasis on equilibrium conditions along perennial rivers, the ability of rare floods to cause substantial change along dryland rivers has led some investigators to describe dryland channels as non- or disequilibrium systems (Graf, 1988). Tooth's (2013) review of dryland rivers, however, emphasized the global diversity of dryland river process and form as a result of varying degrees of aridity, tectonic activity, and structural and lithological settings (Tooth, 2000; Nanson et al., 2002). In addition, the identification of equilibrium or disequilibrium is highly dependent on temporal and spatial frames of reference, so that individual dryland rivers can exhibit both conditions (Tooth and Nanson, 2000).

Isaac Asimov once wrote—perhaps in a moment of frustration—that the only constant is change. This phrase can be adapted to rivers in at least two contexts. First, natural rivers are continually adjusting process and form—spatial distribution of hydraulic forces; water chemistry; sediment entrainment, transport and deposition; instream wood; bed configuration; channel cross-sectional geometry; channel planform; reach gradient and longitudinal profile—in response to changing inputs of water, sediment, and wood, or in response to continuing development of the river. Indeed, one definition of a natural, as opposed to a completely engineered, river is that a natural river possesses geomorphic or physical integrity. Graf (2001) defined *physical integrity* as a set of active fluvial processes and landforms such that the river maintains dynamic equilibrium, with adjustments not exceeding limits of change defined by societal values. River process and form reflect some balance between continually changing external inputs of water, solutes, and sediment and ongoing adjustments within the river. Under these circumstances, considerable insight can also be gained by asking why, under variable external forcing, rivers do not change even more.

The second context for adapting Asimov's phrase to rivers is that, although any river or river segment follows the basic laws of physics and chemistry, characterizing feedbacks between channel process and form using a numerical equation or qualitative conceptual model that applies to all rivers is limited by the place-specific effects of lithology, tectonics, weathering regime, landscape history, river history, the seasonal presence of ice cover or cyclones, and so forth. The only constant within a river network is changes through time and space. The only constant among river networks is river-specific changes in the interactions between process and form. Recognition of these characteristics further emphasizes the importance of understanding landscape context for any particular river network or river segment.

8.5 Rivers with a history

As explained in Chapter 5, a river is a physical system with a history. The influence of previous climatic and tectonic regimes on river form and process can extend back in time beyond the Quaternary because of the slow response of some aspects of river networks to change. Among those characteristics of rivers and drainage basins that can be particularly resistant to changes are:

- topography,
- the spatial arrangement of river channels within a network,
- relief ratio,
- drainage density for river segments larger than first- and second-order channels,
- longitudinal profiles of rivers, and
- valley geometry.

Consequently, inherited characteristics of these features can continue to influence contemporary process and form.

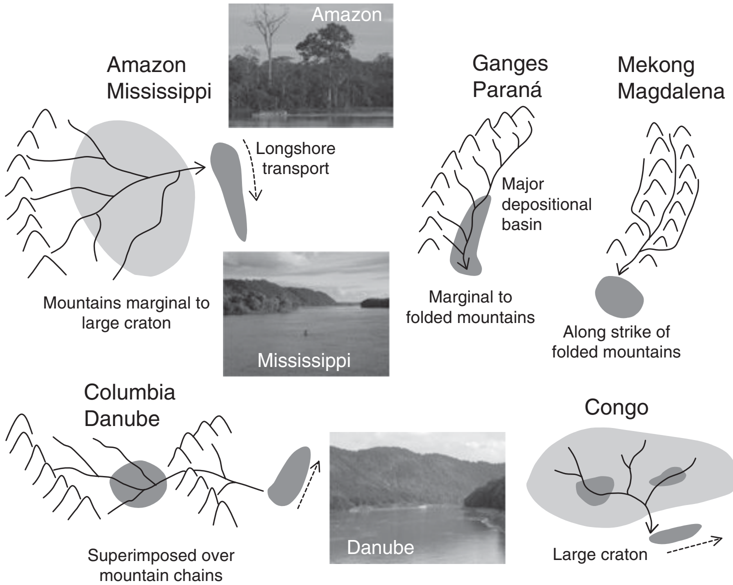

Figure 8.5 Types of big river settings. Lighter gray shading indicates basement bedrock; darker gray shading indicates major depositional areas. (Adapted from Potter 1978, Figure 9.)

An example comes from continental-scale controls on large rivers. The location of large rivers predominantly follows structural lows such as deep-seated rifts, aulacogens (the failed arm of a triple junction in a crustal rift zone), and major fracture systems. Potter (1978) discussed how the continental settings of large rivers fall predominantly into one of the five settings (Figure 8.5). Large rivers draining mountains marginal to a large craton include the Amazon and Mississippi Rivers. Large rivers flowing parallel to folded mountains include the Ganges and Paraná. The Mekong and the Magdalena are large rivers flowing along the strike of folded mountains. Large rivers can be superimposed over mountain chains, as in the case of the Columbia and the Danube, or large rivers can flow across a large craton, as in the case of the Congo.

At smaller spatial scales, a single river channel flowing across different lithologies, structural features, or tectonic zones can exhibit striking differences in river and valley geometry and rate of incision in response to these geological controls that formed millions to hundreds of millions of years ago. Examples include the lower Mississippi River in the eastern United States. This sinuous river can be subdivided into reaches that differ in gradient and sinuosity where the river crosses more erosionally resistant Tertiary-age sediments and fault zones, even though many of these faults show relatively little recent activity (Schumm et al., 2000). Rivers with drainage areas larger than 10 km^2 in the Central Apennines of Italy have long-profile convexities where they cross faults that have undergone an increase in displacement during the past million years, whereas those crossing faults with constant displacement rates lack such convexities (Whittaker et al., 2008).

Landscape configurations or persistent erosional and depositional features relict from Quaternary glaciation provide another example of how past events continue to influence river form and process. Glaciated mountains can have distinctly different valley geometry above and below the elevation limits of Pleistocene valley glaciers, with greater cross-sectional area and steeper valley walls in glacial valleys relative to fluvial valleys (Montgomery, 2002; Amerson et al., 2008). Glaciated and fluvial portions of a mountain range can also be eroding at different rates in response to the effects of differing sediment supply and base level controls (Anderson et al., 2006c). Tributary glacial valleys that eroded to a base level defined by the upper level of the main valley glacier can persist as hanging valleys with large vertical drops between the tributary valley mouth and the main valley floor for thousands of years after glacial ice retreats.

Recessional moraines are persistent depositional features perpendicular to valley orientation that create local base levels and valley segments with lower river gradient and finer substrate than segments immediately up- and downstream, even long after the moraine is incised by a river. Recessional moraines that fill with meltwater as valley glaciers retreat can fail catastrophically. Although individual moraines fail only once, failure of successive moraines along a valley can create numerous outburst floods along a river network over periods of decades to centuries as moraines successively up-valley are abruptly drained. The discharge and stream power of these *outburst floods* typically greatly exceed discharge and stream power generated during "normal" floods induced by rainfall or snowmelt (Cenderelli, 2000; O'Connor et al., 2013), and normal floods may be largely incapable of modifying the outburst-flood erosional and depositional features (Cenderelli and Wohl, 2001). Portions of a valley subject to repeated outburst floods can also become less responsive in that earlier floods have already modified valley morphology to convey exceptionally large discharges (Cenderelli and Wohl, 2003).

The continental-scale ice sheets that covered portions of North America, northern Europe, and northern Asia also left enduring signatures on river networks, diverting existing channels, altering water and sediment supply to channels beyond the ice margins, and changing local river gradients via isostatic flexure of the crust. One of the most spectacular categories of ice sheet effects on river networks is the occurrence of megafloods during periods of glacial retreat. *Megafloods* are relatively short-duration flows that constitute the largest known freshwater floods, with discharges that generally exceed 1 million m^3/s (Baker, 2013). Although other mechanisms, such as failure of rock dams or caldera lake impoundments, can generate megafloods, most were associated with ice-marginal lakes or water released from within the ice sheet. Among the megafloods documented thus far (Baker, 2013) are those of:

- the Channeled Scabland in Washington, USA;
- the Laurentide Ice Sheet ice-marginal lakes in north-central North America, which flowed down the Mississippi, St. Lawrence, Mackenzie, and Hudson Rivers;
- the Patagonian Ice Sheet of southern Argentina and Chile;
- Icelandic jökulhlaups;
- southward drainage from the Fennoscandian Ice Sheet that influenced the English Channel and the North Sea; and
- the northern mountain areas of central Asia, including Kirgizstan, Mongolia, and Siberia.

These exceptionally large floods created erosional and depositional features of such large magnitude and extent that subsequent geomorphic processes during the Holocene have only partially—or in some cases little—modified megaflood terrains.

More recent, Holocene history can also continue to influence river process and form. River response to disturbance partly depends on time elapsed since the last disturbance of a similar magnitude. Investigating erosion along mountainous headwater catchments in which wildfires and subsequent rainfall induced debris flows, Wohl and Pearthree (1991) found that a particular debris flow strongly influenced channel morphology only if a minimum period had passed since the last debris flow. This minimum time interval allowed sufficient sediment to accumulate in the channel and then be eroded by the next debris flow.

Human use of resources can also continue to influence river process and form long after the particular human activity has ceased. Earlier chapters provided numerous examples of such influences, including milldams along rivers in the eastern United States (Walter and Merritts, 2008). Sediment accumulated behind these milldams is now influencing river form and process, as well as downstream sediment and nutrient loads, more than a century after the milldams ceased to be used and maintained. A 1998 study in the southern Appalachian Mountains of the United States found that whole watershed land use in the 1950s was the best predictor of contemporary invertebrate and fish diversity because of persistent effects on aquatic habitat (Harding et al., 1998). Studying streams down which cut logs had been floated for railroad ties in the Medicine Bow National Forest of Wyoming, USA, Young et al. (1994) found that, a century after the log floating had ceased, these streams had less instream wood, lower densities of large riparian trees, lower channel complexity, a greater proportion of riffles, and fewer pools than did otherwise analogous streams that were not used for log floating. As noted earlier, 200 years may be required before instream wood volumes completely recover following timber harvest (Bragg et al., 2000).

The following sections provide a few further, in-depth examples of how the history of geology (lithology, structure, tectonics), climate, biota, and human resource use within a catchment influence contemporary river process and form. These place-specific examples are largely drawn from my own research experience because I can write most effectively of places with which I am familiar. None of the details are unique, however, to these case studies. Many rivers of the semiarid western United States share some aspects of the history of the Upper South Platte River, for example, just as other rivers in the humid tropics share some components of the history of the Rio Chagres. Together, these case studies illustrate the importance of being aware of the greater landscape context when interpreting and managing river process and form.

8.5.1 Upper South Platte River drainage, Colorado, USA

The Upper South Platte River drainage includes numerous tributaries that drain east from the crest of the Colorado Front Range toward the adjacent lowlands of the Great Plains and the Missouri–Mississippi River (Figure S8.6). The Front Range, like other mountain ranges in the region, was generated by compressional tectonics in the late Cretaceous and early Tertiary (Erslev, 2001), and has been relatively quiescent for the past 40 million years (Anderson et al., 2006c). Despite contemporary tectonic stability, the Precambrian-age crystalline rocks that constitute most of the Front Range underwent repeated episodes of uplift and erosion following the Precambrian, and are now densely jointed. Precambrian shear zones that were reactivated during the most recent orogeny create particularly densely jointed bedrock that correlates with the location of wider valley segments with lower downstream gradients, and the creation and preservation of strath terraces (Ehlen and Wohl, 2002; Wohl, 2008a).

Exhumation of the adjacent, sediment-filled basins beyond the mountain front began sometime during the late Cenozoic, and many of the basins have now been eroded to depths of 1 km or greater (Leonard, 2002). Basin exhumation lowers base level for the rivers draining the Front Range, creating knickpoints or knickzones that currently lie at varying distances upstream from the mountain front, partly as a function of differing drainage area, discharge, and incisional capabilities on the rivers tributary to the South Platte (Anderson et al., 2006c).

The upper elevations of most valleys were also influenced by Pleistocene valley glaciers (Madole et al., 1998). In addition to eroding bedrock valley boundaries and creating knickpoints along tributary valleys, the glaciers deposited large moraines that continue to create local base levels for upstream valley segments (Wohl et al., 2004). Wide, low gradient valley segments immediately upstream from the terminal moraine are likely to host beaver meadows where historically abundant beaver populations created numerous dams, multi-thread channel networks, and thick sequences of organic-rich fine sediments (Westbrook et al., 2006, 2011; Kramer et al., 2012; Polvi and Wohl, 2012; Wohl et al., 2012c).

The lower limit of Pleistocene glaciation, ~2300 m elevation, corresponds to a transition in modern hydroclimatology. Above this elevation, snowmelt dominates the annual streamflow regime and discharge per unit drainage area seldom exceeds 1.1 $m^3/s/km^2$ (Jarrett, 1989). Below this elevation, snowmelt continues to be the primary source of annual peak flow, but late summer convective storms can produce localized flash flooding that generates as much as 40 $m^3/s/km^2$ of discharge (Jarrett, 1989). These flash floods generate sufficiently high values of shear stress and stream power to create substantial and persistent erosion in deep, narrow valley segments and deposition in wide, lower-gradient segments (Shroba et al., 1979). Floods with a recurrence interval of hundreds of years thus create the valley-bottom template at middle to lower elevations in the Front Range.

Hillslope disturbance regimes also influence valley-bottom process and form. Primary sources of hillslope instability are intense precipitation at elevations below 2300 m, or disruption of the forest cover by wildfire at all elevations. Steppe vegetation at the base of the mountains gives way to montane forests (1830–2740 m elevation), with open ponderosa pine (*Pinus ponderosa*) woodlands in the lower montane (1830–2350 m) and slightly more dense mixed conifer forests that include Douglas fir (*Pseudotsuga menziesii*) in the upper montane (2440–2740 m). Frequent, low-severity fires that burn mainly the ground surface over areas of approximately 100 ha recur at intervals of 5–30 years in the lower montane zone (Veblen and Donnegan, 2005). Because the canopy and the root structure of the forest remain intact, these fires are unlikely to trigger widespread hillslope instability, although destruction of the surface organic layer can create local slopewash and rilling.

A complex pattern of mixed low- and high-severity fires that burn areas of approximately 100 ha and recur at intervals of 40–100 years occurs in the middle and the upper montane zone. High-severity fires can trigger widespread hillslope instability when unconsolidated hillslope sediment is exposed to snowmelt, rainfall, and dry ravel.

The subalpine forest (2740–3400 m) includes Engelmann spruce (*Picea engelmannii*), subalpine fir (*Abies lasiocarpa*), lodgepole pine (*Pinus contorta*), aspen (*Populus tremuloides*) and other species. The subalpine forest is characterized by infrequent, high-severity fires that kill all canopy trees over areas of hundreds to thousands of hectares at intervals greater than 100 years (Veblen and Donnegan, 2005). As at lower elevations, these fires can trigger hillslope instability.

People have lived in the Front Range for at least 12,000 years (Benedict, 1992), but the hunter–gatherer cultures of the early inhabitants do not appear to have altered river process and form. Human impacts to rivers started with the advent of beaver trapping for furs during the first decades of the nineteenth century, and accelerated with the 1859 discovery of placer gold in several tributaries of the Upper South Platte River (Figure S8.7), and the associated deforestation, construction of roads and railroads (Figure S8.8), floating of cut logs downstream to processing areas (Figure S8.9), flow regulation for agricultural water supply (Figure S8.10), and urbanization (Wohl, 2001). Although a few relatively unimpacted, reference rivers remain in the Front Range, process and form have changed substantially during the past two centuries along the majority of rivers. This can be illustrated by considering the effects of instream wood.

River segments flowing through old-growth forest, and which experienced minimal historical disruption from activities such as placer mining, tend to have greater volumes of instream wood per unit channel area, and larger and more closely spaced channel-spanning logjams (Beckman, 2013; Wohl and Beckman, 2014). Larger instream wood loads correspond to more physical diversity of channel planform along the length of a river (Wohl, 2011c; Polvi and Wohl, 2013), greater storage of nutrients (Wohl et al., 2012c), and greater abundance and diversity of riparian and aquatic habitat (Richmond and Fausch, 1995). In the absence of wood, channels tend to be much more uniform with respect to cross-sectional geometry (e.g., fewer pools, and a more continuous riffle-run morphology), planform (almost entirely single-thread channels, sometimes with lower sinuosity), and aquatic and riparian habitat (Figure S8.11). Water, dissolved and particulate nutrients, and sediment all move more rapidly downstream in the absence of obstructions and in-channel storage associated with instream wood, and channel–floodplain connectivity and hyporheic exchange decrease. Pronounced longitudinal variations in valley geometry associated with

Precambrian shear zones and Pleistocene glaciations strongly influence wood transport and storage at channel lengths of 10^1–10^2 m (Wohl and Cadol, 2011), and wildfire and other forest disturbances influence wood recruitment to rivers (Wohl, 2011b), but the historical legacy of nineteenth and twentieth century human activities along these rivers dominates instream wood loads and, consequently, river process and form.

Effective management and restoration of rivers in the Upper South Platte drainage must recognize at least four salient factors:

- substantial reach-scale (10^1–10^2 m) variation in process and form imposed by geologic factors (Figure S8.2), which means that many river characteristics are best described using the conceptual model of process domains;
- episodic natural disturbances, primarily in the form of debris flows and floods associated with wildfires and/or convective storms, which create persistent erosional and depositional features that shape valley and channel geometry at time spans of 10^1–10^3 years;
- simplification and homogenization of most rivers as a result of nineteenth and twentieth century human activities, several of which (e.g., flow regulation) continue today; and
- ongoing climate change toward a warmer, drier climate with smaller annual snowpack, earlier snowmelt, changes in forest composition associated with changing climate and disturbance (wildfire, insects, blowdowns), and increased human consumptive demands for surface waters.

Under these circumstances, maintaining and restoring the physical and ecological integrity of rivers in the Upper South Platte River drainage presents a substantial challenge.

8.5.2 Upper Rio Chagres, Panama

The Upper Rio Chagres is the largest headwater basin within the 2982 km^2 watershed of the Panama Canal, and supplies almost half of the water needed for the operation of the canal (Harmon, 2005a). Consequently, the Upper Rio Chagres has received more scientific attention than many of the watersheds in the neotropics of Central America.

Panama is an east-west-oriented isthmus between North and South America composed of a diverse suite of geological units created and assembled since the late Cretaceous (Harmon, 2005b). The igneous rocks that comprise most of Panama formed during the Tertiary as an oceanic plateau and volcanic island arc complex. This arc began to collide with northwestern South America approximately 10 million years ago and the land bridge between the two continents was in place around 3 million years ago (Harmon, 2005b). At present, the Pacific side of Panama is a geologically active margin with a deep oceanic trench, narrow marine shelf, active subduction, volcanic activity, and earthquakes, whereas the Atlantic side is a passive, stable margin (Harmon, 2005b). The bedrock of the Upper Rio Chagres basin is mainly a mixture of volcanic and intrusive rocks that range from felsic to ultramafic, most of which are strongly deformed and chemically altered (Wörner et al., 2005).

The Rio Chagres basin was divided into an upper and a lower basin by completion of the Panama Canal and associated dams in 1914, which flooded a significant portion of the original Chagres basin (Harmon, 2005a) (Figure S8.12). The 1936 construction of Madden Dam formed Lake Alhajuela: the Upper Rio Chagres drains 414 km^2 above this lake. The river network of the Upper Chagres is deeply incised into an extremely steep terrain. Hillslopes exceed 45 degrees in over 90% of basin, most of which is covered by intact primary tropical and seasonally tropical rainforest. Trees can attain a height of 30 m and a diameter of 2.2 m (Wohl, 2005). Mean annual precipitation at the basin outlet is 2590 mm, 90% of which falls during May to December. Rainfall in the upper watershed is unmeasured, but likely closer to 5000 mm based on vegetation type.

Soil surfaces are relatively dry and water repellent at the start of the wet season, and soil tension cracks have developed. Landslides and tree falls create persistent, uneven hillslope topography and numerous small surface depressions. These depressions accumulate water relatively quickly during rains at

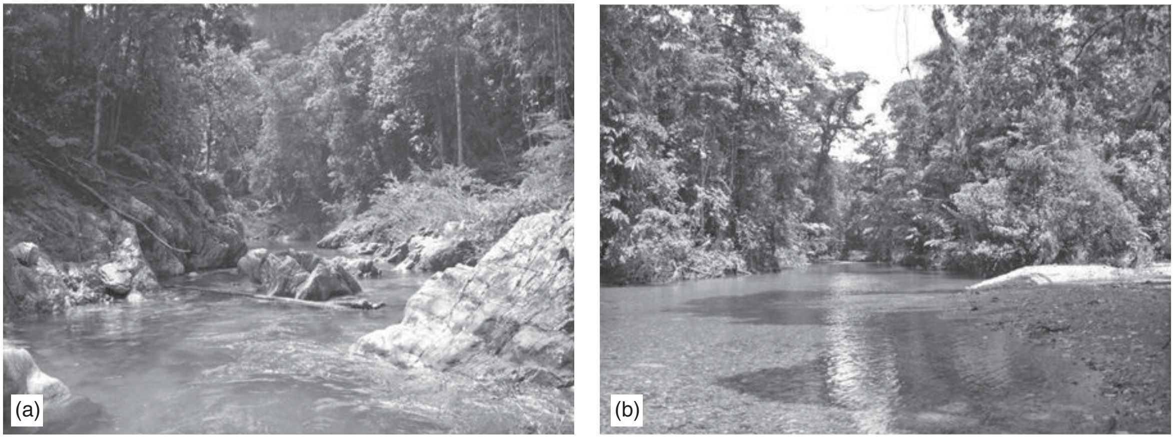

Figure 8.6 Views of diverse segments of Upper Chagres River formed in (a) mafic bedrock (gradient 0.014, channel width 30 m), and (b) felsic rocks (gradient 0.001, channel width 35 m).

the start of the wet season, and these rains produce anomalously high runoff volumes. Water repellent soils limit infiltration from the depressions, which, along with the soil cracks, feed a well-developed network of soil pipes that rapidly deliver runoff to stream channels. As the wet season continues and the soil surface becomes moist, water repellency and soil cracks disappear, infiltration into the soil matrix fills the soil water storage reservoir, less runoff moves downslope via soil pipes, and runoff volume per precipitation input decreases. Runoff volume again increases later in the wet season, however, as saturated soils create saturated overland flow (Hendrickx et al., 2005). Runoff efficiency of a storm—defined as the ratio of the volume of runoff to the volume of rainfall—varies from 6% to 59% during the wet season in the Upper Chagres, with an average of 40% (Niedzialek and Ogden, 2005).

Sequential changes in hillslope hydrology during the wet season are reflected in streamflow. The increase in saturation overland flow later in the wet season, for example, appears as increasingly higher quasi-stable base flows (Niedzialek and Ogden, 2005). Unit discharges in the headwaters portion of the basin can be quite large: heavy rainfall in 2007 resulted in a peak discharge of 41 $m^3/s/km^2$ in a 20.6 km^2 catchment (Wohl and Ogden, 2013).

Bedrock is discontinuously exposed along the bed and banks of channels throughout the Upper Chagres basin, and is likely never more than 2–3 m below the channel. Mafic units are the most resistant to weathering and erosion, and correspond to steep, narrow valley segments and knickpoints (Wohl and Springer, 2005). Granitic rocks are more readily weathered and correspond to straighter, wider valley segments (Wörner et al., 2005) (Figure 8.6). Downstream hydraulic geometry relations are well developed despite these local variations, with exponents similar to average values for rivers worldwide (Wohl, 2005) (Figure S8.13). This suggests that high values of discharge per unit drainage area generate sufficient erosional force to reduce variations in channel geometry associated with local differences in rock erodibility, although poor correlations between grain size or reach gradient and drainage area (Rengers and Wohl, 2007) suggest that landslides and bedrock lithology strongly influence some aspects of river process and form.

Landslides are frequent and widespread throughout the Upper Chagres (Figure S8.14), and are an important mechanism of instream wood recruitment. Wood recruitment and retention are extremely spatially and temporally episodic along the Upper Chagres. High decay rates of wood and large fluvial transport capacity combine to create extremely low "background" levels of instream wood. When intense rainfall generates landslides, however, large volumes of wood can be abruptly introduced to the river network. Widespread landsliding and wood recruitment can result in the formation of channel-spanning logjams and upstream wedges of sediment accumulation in tributary channels (Figure S8.15) and at sites of reduced transport in the main channel, such as bends or expansions (Wohl et al., 2009). These jams persist for only 1–2 years, although longitudinally discontinuous fill terraces 1–2 m above the active channel may persist when a jam breaks up and incision removes sediment stored above the jam along the channel axis (Wohl et al., 2009). Infrequent storms that generate greater rainfall over much of the upper basin trigger even more abundant landslides and wood recruitment, but sustained high flows limit formation of logjams and instead flush the newly recruited wood all the way through the Upper Chagres river network and into Lake Alhajuela (Wohl and Ogden, 2013). This represents a substantial flux of organic carbon from the basin. Wohl and Ogden (2013) estimated values of 24 Mg C/km^2, which is an order of magnitude higher than background rates of wood-based carbon export from other catchments.

In situ-produced cosmogenic ^{10}Be analyses indicate sediment generation rates averaging ~270 tons/km/year in the Upper Chagres, a value comparable to other regions with high rainfall and runoff (Nichols et al., 2005). Adjacent drainages with extensive land clearance have twofold to threefold greater sediment yields per unit drainage area (Nichols et al., 2005). Land clearance represents the greatest immediate potential source of changes in sediment and water yield, water quality, and carbon stocks within the Upper Chagres (Dale et al., 2005). Although much of the watershed lies within the Parque Nacional Chagres, illegal clearing and settlement within the basin have increased during the past decade as human population increases rapidly within the region.

As in the Upper South Platte drainage, effective management and restoration of rivers in the Upper Chagres drainage basin requires recognition of fundamental river processes governed by the geologic and climatic setting. This setting creates

- high rainfall, runoff, and discharge per unit area,
- widespread, frequent landslides and introduction of sediment, wood, and organic carbon to rivers,
- high transport capacity within rivers and attendant rapid changes in erosional and depositional features along channels, and
- a river network that is likely to be highly responsive to changes in land cover and climate.

8.5.3 Mackenzie River drainage, Canada

The Mackenzie River drains 1.8 million km^2 in Canada and is the fourth largest Arctic river with respect to water discharge and the largest in terms of sediment discharge (Holmes et al., 2002; Rachold et al., 2004). The mainstem Mackenzie flows north to the Arctic Ocean from Great Slave Lake, but primary tributaries extend much farther south (Figure S8.16). Permafrost underlies two-thirds of the Mackenzie drainage (Heginbottom, 2000), and the presence of this permanently frozen ground strongly influences hydrology and geomorphology within the region. The drainage is unique in containing 8 of the 15 ecozones identified in Canada (de Rham et al., 2008a), but much of the drainage basin (79%) is covered by forest, and the river has long been recognized as a primary source of wood export to the Arctic Ocean (Eggertsson, 1994). Like other high-latitude, forested catchments, the Mackenzie also stores substantial carbon in forest and wetland ecosystems and associated peat deposits (Dixon et al., 1994).

The western Mackenzie drainage includes individual ranges of the Canadian Rockies, whereas the central and eastern portions of the basin are on the low-relief Canadian Shield and Interior Plains (Woo and Rouse, 2008). The western mountainous terrain and the central and eastern plains have distinctly different pathways for water and sediment moving from uplands and into the river network.

The lowlands of the Mackenzie drainage are an arid to semiarid region that receives relatively little precipitation, but nonetheless contains tens of thousands of small lakes (Figure 8.7). This apparent paradox reflects low evaporation rates and, for very

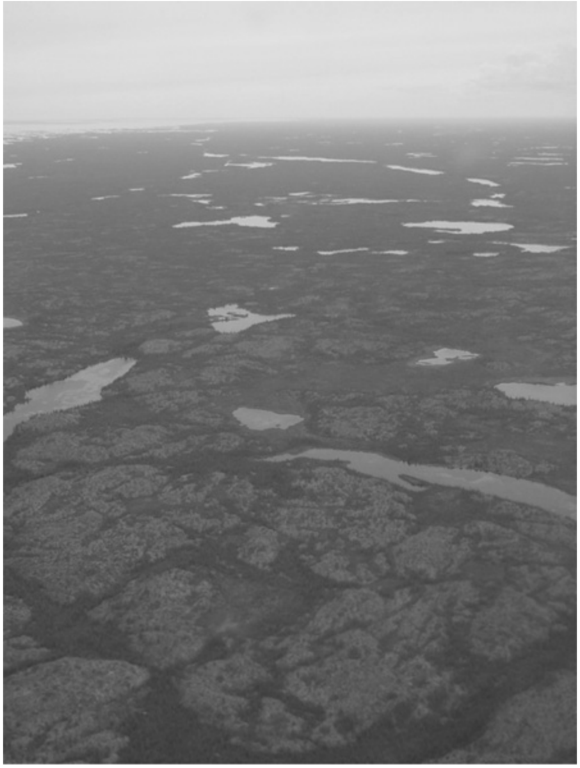

Figure 8.7 An aerial view of widely scattered small lakes, bedrock outcrops, and patches of forest in the lowlands of the eastern Mackenzie River drainage basin.

shallow lakes, impermeable permafrost that prevents infiltration of precipitation or meltwater when soil ice thaws during summer.

The element threshold concept described in Section 2.2.1 was developed for the lowlands of the Mackenzie drainage. The central tenet of this concept is that spatially and hydrologically distinct areas such as bedrock uplands, soil-filled valleys, and lakes create a mosaic on the landscape. Each area, or element, within the mosaic can store, contribute, or transmit water, and the properties of each element are not necessarily synchronized with those of adjacent elements, so that a bedrock upland can be contributing water while an adjacent lake is storing water, and an upslope element may contribute overland flow before a downslope element (Spence and Woo, 2006). The lowlands contribute sediment to the river network primarily through bank erosion.

The mountainous uplands also have extensive permafrost, but the steeper terrain likely contributes to a runoff pattern more like the source area concept (Section 2.2.1). On a per unit area basis, the mountainous areas contribute more runoff than the lowlands (Woo and Thorne, 2003). The uplands also contribute substantial sediment to the mainstem, and much of this sediment enters the river network via landslides (Figure S8.17). A 1990s inventory of 3400 landslides within the Mackenzie drainage found that 69% of the landslides occurred in unconsolidated sediments. Landslides are concentrated along the banks of the Mackenzie and its tributary channels where the rivers have eroded Quaternary sediments and along steep slopes in the mountains west of the Mackenzie River (Aylsworth et al., 2000).

Runoff in the Mackenzie drainage is dominated by high spring flows, but the presence of seasonal ice creates unique stage-discharge relations (Hicks and Beltaos, 2008). The additional hydraulic resistance of a stable ice cover elevates water levels in a channel (Prowse and Beltaos, 2002), and this effect is greatest when the ice cover is hydraulically most rough, such as during freeze-over and break-up. Pronounced seasonal changes in hydrographs also reflect storage of significant quantities of water as ice during freeze-over and subsequent release during break-up (Prowse and Ferrick, 2002). Early winter freeze-over can create the lowest flow of the year, even though actual runoff production reaches a minimum later in the winter. This apparent disconnect results from three factors: reduced contributing area because anchor ice frozen to the river bed cuts off groundwater inflows; storage of water within ice; and backwater storage upstream from ice-covered river segments in which the ice reduces the channel cross-sectional area (Prowse and Beltaos, 2002). Release of stored water and melt of the ice cover can significantly contribute to spring peak flows (Prowse and Carter, 2002). Peak water levels can be enhanced by ice jamming, which can also cause accelerated channel and bank erosion (Goulding et al., 2009a; Ettema and Kempema, 2012).

Peak flows can create extensive overbank flooding in portions of the drainage such as the Mackenzie Delta, where overbank flows strongly influence hydrologic and biogeochemical exchanges between channels and delta lakes (Goulding et al., 2009b; Prowse et al., 2011). The delta covers 12,000 km^2 and contains more than 25,000 lakes and a sinuous maze of distributary channels (Goulding et al., 2009a, 2009b). Mean annual flow through the delta is 8300 m^3/s, but peak flows can reach 14,000 m^3/s during May and June (Mackay, 1963). Hydraulic forces, along with transport of ice blocks (Goulding et al., 2009a, 2009b) and wood, facilitate constant bank erosion (Figure S8.18) and sedimentation. An estimated 128 million tons of sediment reach the delta annually from the Mackenzie River and the tributary Peel River (Carson et al., 1998). Approximately 103 million tons are at least temporarily deposited in the delta, of which 43 million tons remain: the rest moves from the delta into Mackenzie Bay and the Beaufort Sea (Carson et al., 1998). The main distributary channels of the delta have nonetheless remained relatively stable since they were mapped by the Franklin expedition in 1826, which may reflect the stabilizing presence of permafrost. As the permafrost underlying the region melts, rates of channel avulsion, bank erosion, and wood recruitment have the potential to increase dramatically.

The Mackenzie drainage is at the forefront of environmental changes associated with global warming, as are other rivers draining to the Arctic (Guo et al., 2007). Increases in average air temperature will cascade through arctic ecosystems in complex ways, altering the distribution of permafrost, surface water hydrology (especially frequency and magnitude of ice jam floods), vegetation, and biogeochemical fluxes (Francis et al., 2009a; Frey and McClelland, 2009). As permafrost recedes, thawing of frozen organic soils and stream banks will release carbon into rivers via groundwater pathways, surface erosion, bank instability, and landslides (Dyke et al., 1997; Holmes et al., 2008). Warming air temperatures will not only influence precipitation and runoff, but will also alter the dynamics of river ice by delaying freeze-over and changing the timing of break-up (Prowse and Beltaos, 2002). Records during the period 1974–2006 indicate a trend toward longer melt interval prior to break-up, lower peak discharge, slower rate of rise in discharge, and lesser ice thickness in the Mackenzie Delta (Goulding et al., 2009a). Analysis of break-up during the period 1913–2002 also indicates significantly earlier trends in the timing of break-up in upstream portions of major tributaries of the Mackenzie drainage (de Rham et al., 2008b).

Despite the low population density and remote location of most of the Mackenzie drainage, we have a relatively extensive knowledge base for this river network partly because of the history of proposed and actual resource extraction. Starting with a proposed natural gas pipeline during the 1970s that led to a range of baseline environmental studies, and continuing with additional proposals for oil and gas extraction and transmission corridors at present, we know more about climate, hydrology, biogeochemistry, forest ecology, and geomorphology in the Mackenzie drainage than in other Arctic drainages such as the Yukon in North America and the major rivers of Siberia. Effective management of rivers in the Mackenzie drainage requires recognition of distinctive factors associated with the geologic and climatic setting, including

- permafrost that limits infiltration while intact,
- seasonal ice cover that alters flow resistance, water storage, downstream transmission of runoff during snowmelt, and timing and magnitude of peak flows,
- differences in runoff and sediment yields from mountainous versus lowland portions of the drainage, and
- rapid changes in all components of the hydrologic cycle as global air temperatures rise.

8.5.4 Oregon Coast Range, USA

The Oregon Coast Range is a rugged, deeply dissected mountain range in the northwestern United States (Figure S8.19). The range sits over the Cascadia subduction zone and has experienced variable uplift rates during the past 20–30 million years. Rock uplift rates average 30–300 m per million years, basin denudation rates average 50–80 m per million years, and bedrock lowering averages 70 m per million years (Heimsath et al., 2001). The range is underlain by a thick sequence of arkosic sandstone and siltstone, with minor areas of mafic volcanic and intrusive rocks (Personius et al., 1993). Summits in the Oregon Coast Range lie below 1250 m elevation, and the range was not glaciated during the Pleistocene. Peaks and ridges are soil mantled, with a typical profile of colluvium overlying saprolite, fractured bedrock, and then intact bedrock that can be 1.5 m thick (Heimsath et al., 2001).

The combination of active uplift, readily erodible sedimentary rocks, and wet climate (mean annual precipitation averages 200 cm) creates abundant landslides and debris flows. The Coast Range is densely forested, and landslides occur in areas of reduced root strength associated with diverse vegetation (Roering et al., 2003). Landslides also correlate with greater amounts of siltstone relative to sandstone, and occur preferentially on slopes on which the downslope aspect corresponds to the direction of bedrock dip (Roering et al., 2005). Wildfires are also important in destabilizing slopes. Post-fire erosion rates exceed long-term rates (~0.1 mm/year) by a factor of six, and numerical simulations suggest that fire-related processes may account for ~50% of sediment production on steep slopes (Roering and Gerber, 2005).

Rivers in the Oregon Coast Range flow mostly on bedrock or thin alluvial beds bordered by bedrock, and bedrock straths and discontinuous terraces are widespread (Personius et al., 1993). An episode of regional aggradation at the start of the Holocene has been interpreted to reflect climate-induced changes in the frequency of evacuation of colluvium from hillslope hollows (Personius et al., 1993) and increased fire frequency and sediment yield (Roering and Gerber, 2005).

Spatial variations in the gradient and longitudinal profile of rivers in the Oregon Coast Range have been used to infer the location of synclinal tilting and uplift (Rhea, 1993). A zone approximately 20 km wide in which channel gradients are about twice the regional average coincides with the strike of N-S-trending folds, and likely reflects differential rock uplift (Kobor and Roering, 2004). Colluvial hollows within this zone are also steeper than in adjacent areas and may experience more frequent landslides (Kobor and Roering, 2004).

Landslides and debris flows strongly influence river form and process in the Oregon Coast Range. Terrain prone to large landslides has lower values of drainage density (Roering et al., 2005). Debris-flow recurrence intervals vary from ~100 to 400 years in headwater basins (May and Gresswell, 2004). Once debris flows enter the river network, they typically strip the steeper channel segments to bedrock, evacuating stored sediment and instream wood. These channel segments can remain bedrock channels for substantial periods of time, but wood that accumulates in the channel can effectively trap sediment, and low-order channels with wood can form one of the more important sediment storage reservoirs in this steep landscape (May and Gresswell, 2003). Tributary basins with larger drainage areas have more potential landslide source area and a greater frequency of scouring debris flows than smaller basins, facilitating the formation of larger, more persistent debris fans at channel confluences (May and Gresswell, 2004).

Debris flows commonly dam headwater valleys with deposits of sediment and wood, causing upstream alluviation that covers bedrock channel segments (Lancaster and Grant, 2006). These temporary sediment storage sites are evacuated at rates that reflect process transitions. Channel segments higher in the river network that experience frequent debris-flow erosion have shorter transit time for sediment than segments lower in the network that experience only fluvial erosion (Lancaster and Casebeer, 2007). Although most sediment in both portions of the network has relatively short transit times (<600 years), significant volumes of sediment can remain for thousands of years, so that these valley-bottom sediment reservoirs buffer downstream aquatic habitat from hillslope disturbance (Lancaster and Casebeer, 2007).

The Oregon Coast Range has also been an area of intensive timber harvest for more than a century. As noted in Chapter 7, timber harvest typically results in substantial changes to water and sediment yield, as well as instream wood recruitment. In the Coast

Range, instream wood forms about half of the pools in old-growth forests, but fewer pools in industrial forests, which have substantially fewer jams per length of channel (Montgomery et al., 2003b). Logjams that do form within river networks in industrial forests tend to be associated with debris flows and to form at river confluences, in contrast to the more widely distributed logjams of old-growth forests. The widely distributed jams in old-growth river networks facilitate storage of channel alluvium and thus dampen bedrock incision rates. Jams in old-growth areas also contribute to the presence of three process domains (Montgomery et al., 2003b). (1) An alluvial zone exists in the lower portions of mainstream channels where channel and valley gradients are sufficiently low to permit a persistent sediment layer over the bedrock. (2) A forced alluvial zone is present in the middle reaches of the river network, where instream wood increases flow resistance and creates backwater effects that allow sediment cover to persist along valley segments that would otherwise have exposed bedrock in the streambed. (3) A zone dominated by debris flows occurs in headwater channels with drainage areas less than ~1 km and channel gradients >0.2.

In the Oregon Coast Range, identifying the factors that correlate with greater landslide activity facilitates identification of the areas of greatest landscape change, as well as areas likely to be most sensitive to climate variability and land use. Effective management and restoration of rivers in the Coast Range requires recognition of

- the importance of tectonics, lithology, and climate in promoting hillslope instability,
- the resulting spatial and temporal variability in sediment supplied to rivers,
- patterns of sediment storage and evacuation within river networks that reflect process domains, debris flow inputs, and instream wood dynamics, and
- the influence of land use, specifically timber harvest, on slope stability and wood recruitment to channels, and the associated differences in river form and process between undisturbed and industrial forests.

8.5.5 Yuma Wash, Arizona, USA

Yuma Wash is a mixed sand- and gravel-bed channel that drains 186 km^2 in southwestern Arizona, an extremely arid region (Figure S8.20). The channel is formed primarily in alluvium, but bedrock outcrops discontinuously along the upper reaches of the channel network (Figure S8.21). Headwater reaches are relatively narrow, single-thread channels incised into bedrock or cohesive alluvium cemented by secondary carbonate. Further downstream, Yuma Wash grows progressively wider, until valley width exceeds 450 m and the channel becomes braided. The outer edges of the braided channel pattern are defined by Holocene and Pleistocene terraces with steep risers of indurated alluvium.

Vegetation cover in the surrounding uplands is only 1%–5%, but averages 31% within the middle and lower portions of the channel network (Merritt and Wohl, 2003). Dominant vegetation includes woody xeric species, succulents and grasses (Merritt and Wohl, 2003). Clumps of vegetation form linear or lemniscate patterns on depositional surfaces in the valley bottoms. Woody species growing on bars and adjacent to active channels are resistant to removal by scour and are adapted to survive periodic flooding (Figure S8.22).

The mainstem and all of the tributaries are ephemeral, and flows occur less frequently than once a year. The region receives a total annual average rainfall of 93 mm (NOAA, 1998) from convective thunderstorms, frontal systems, and dissipating tropical storms. Most of the summer rainfall comes from isolated, fast-moving convective storms that cause localized flash flooding. Frontal storms during autumn and winter create more widespread but typically low-intensity rainfall that generates little stream flow. Floods with the longest duration (typically hours to a few days) and largest spatial extent result from dissipating tropical storms during the autumn months.

Examination of channel change resulting from 1997 Hurricane Nora indicated that wider, braided reaches aggraded substantially during the 1997 flood, whereas narrower reaches incised. Patterns of aggradation and incision reflected a threshold relationship between flow depth and flow resistance associated with vegetated bars. Degradation occurred as long as flow was confined within a channel or subchannel. At flow depths sufficiently high to overtop vegetated bars, greater flow resistance associated with the vegetation facilitated lower velocities, deposition, and aggradation (Merritt and Wohl, 2003).

Downslope from the steep, mountainous terrain that characterizes the headwaters of Yuma Wash, channels are incised into more gently sloping alluvial piedmont surfaces of alluvial fans and pediments (Figure S8.23). Eolian sedimentation and accumulation of silt facilitate the formation of extensive desert pavement (Bacon et al., 2010). Desert pavement consists of a coarse upper layer of pebble to cobble size clasts that are typically covered in a secondary iron–manganese weathering rind known as desert varnish. Many of the upland surfaces host cryptobiotic soil crusts that help to stabilize finer sediment near the surface. Desert varnish, desert pavement, and cryptobiotic soil crusts take many years to form, but these surfaces are disrupted by military activities in the vicinity of Yuma Wash.

Yuma Wash lies within the US Army's Yuma Proving Ground (YPG), which was established in 1943 as a hot desert training center. YPG is used primarily for tactical ground exercises, testing of heavy artillery, and aerial release of heavy equipment. Starting during the era of World War II, tactical ground exercises focused on the piedmont uplands. Disturbance of desert pavement and cryptobiotic soils in association with these activities has proven to be persistent: the outlines of roads and tent camps from that era still appear clearly in contemporary aerial imagery. Consequently, military exercises have more recently focused in the channel network. Most active roads through the area, for example, use dry washes. Although dry washes might be thought to have lesser biological value than intermittent or perennial channels, the vegetative abundance and diversity of the channel network at Yuma Wash, as well as the importance of woody riparian vegetation in creating flow resistance and limiting channel erosion, suggest the potential for substantial ecological losses and reduced channel stability as a result of military activities in the channel network.

Ongoing research designed to identify the most ecologically diverse and resilient channel segments focuses on a six-part channel classification (Sutfin, 2013) of

- montane bedrock channels entirely confined by exposed bedrock and lacking persistent alluvium;
- upper piedmont bedrock with alluvium channels that are at least partly confined by bedrock but contain enough alluvium to create bedforms that persist through time;
- incised alluvium channels bounded only by unconsolidated alluvium into which the channel segment is incised;
- depositional braided washes with multi-thread channel planform regardless of the degree and composition of confining material;
- piedmont headwater channels that are first- or second-order streams confined only by unconsolidated alluvium and which initiate as secondary channels on piedmont surfaces; and
- flood-outs, a name originally applied in Australia (Tooth, 1999) to channels in arid regions in which all signs of defined channels and surface flow disappear as a result of transmission losses (evaporation and infiltration).

Statistical analyses indicate that these six stream types differ significantly in terms of stream gradient, width/depth ratio, entrenchment ratio (ratio of valley width at two times the reference stage (reference stage here is stage of a relatively common discharge analogous to bankfull)), shear stress, and stream power per unit area (Sutfin, 2013).

Effective channel management and restoration in Yuma Wash requires recognizing some of the unique aspects of this hot desert environment, including

- the temporally and spatially discontinuous nature of channel change—individual channel segments within the river network commonly do not change in the same direction (i.e., some reaches are incising while others are aggrading), at the same time, or at the same rate,

- not all rainfall produces runoff or stream flow, and only the most extensive and sustained rainfall above a minimum intensity threshold produces discharge through substantial portions of the channel network—this typically does not happen every year,
- discharge can increase downstream to some extent, but at some point in the network discharge during a single flood decreases downstream as a result of infiltration and evaporation losses,
- alluvial upland surfaces are highly sensitive to physical disruptions of desert pavement and cryptobiotic soil crusts, and disruption of these features likely alters rainfall–runoff relations, and
- woody riparian vegetation strongly influences flow resistance and channel adjustment during floods.

8.6 The greater context

The nineteenth-century American conservationist John Muir famously wrote in 1911, "When we try to pick out anything by itself, we find it hitched to everything else in the Universe." Many other thinkers have stated the same concept in different words, recognizing the interconnectedness of natural systems and people. Rivers are no exception to this rule, as I have emphasized from the opening pages in this book. As the intensity and extent of human alteration of river form and process have accelerated globally since the mid-twentieth century, abundant evidence has appeared that reflects riverine influences on the entire critical zone. Changes in coastal environments provide a vivid example.

Human activities now dominate nitrogen budgets in regions such as Asia, Europe, and North America (Boyer et al., 2006). Substantial increases in nitrogen yields to rivers result from industrial-scale agriculture, feedlots, and septic systems. Riverine corridors have been simplified via channelization, removal of riparian vegetation, levees and flow regulation, all of which reduce channel–floodplain connectivity. This simplification reduces nitrogen retention and processing by rivers. Greater inputs and less storage create a "one-two punch," resulting in substantially increased nitrogen fluxes down rivers to coastal areas. This has created eutrophication of estuaries and other nearshore environments.

For the most part, we cannot yet predict in detail how diverse changes in rivers resulting directly and indirectly from human activities will affect the greater landscape. Synthesizing studies in the northern high latitudes, for example, Woo (2010) explained how freshwater discharge during spring snowmelt influences numerous and diverse processes in the nearshore zone and greater Arctic Ocean (Figure 8.8). Freshwater discharge influences the dynamics of coastal sea ice, terrestrial sediment and organic matter plumes into the ocean, and

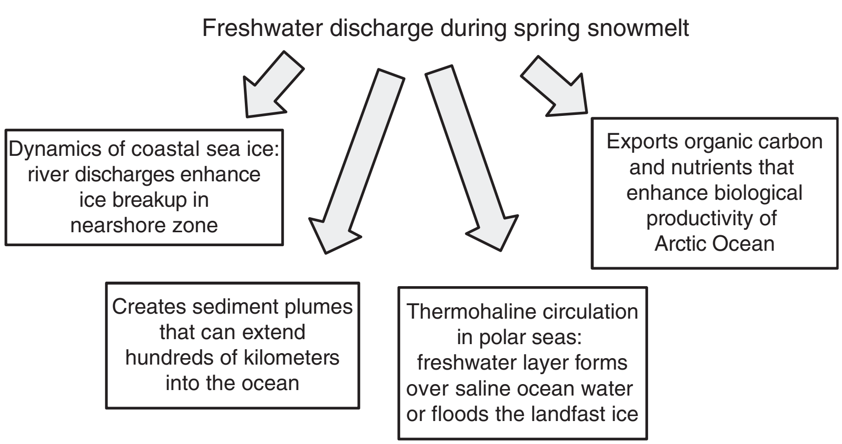

Figure 8.8 Schematic diagram of the effects of freshwater discharge during spring snowmelt from rivers draining to the Arctic Ocean.

thermohaline circulation in polar seas. The complex effects of global warming on these interactions remain largely unknown. Less sea ice may modify existing energy and moisture fluxes, for example, and thus alter coastal storm patterns and inland water balances (Woo, 2010).

Unfortunately, we too commonly recognize the importance of landscape context and connectivity within and between river networks once our activities have altered connectivity and caused unintended negative consequences. Examples span spatial scales from headwaters to the world's largest rivers, climatic zones from hot and cold deserts to rainforests, and tectonically active and passive terrains with varying lithology and structure. The following list expands on some of the examples mentioned briefly in Chapter 1 and introduces others.

- Nineteenth- and twentieth-century wetland drainage and channelization along thousands of kilometers of small, headwater streams in the Illinois River basin of the central United States reduced channel–floodplain connectivity and caused precipitous declines in fish and waterbird communities in this once extraordinarily productive watershed (Mattingly et al., 1993). Here, reduced lateral connectivity alters biotic communities.
- The Los Angeles River drains 2200 km^2 of southern California, USA, flowing <60 km from the Santa Susana Mountains to the Pacific Ocean. The fire-prone chaparral vegetation in these mountains and the dry climate historically contributed substantial amounts of sediment to the nearshore zone via the river: on the order of 600,000 m^3 per year (Flick, 1993). The combined effects of sediment detention basins in the high-relief headwaters and extensive channel stabilization along the river's course (tributaries are confined by concrete within 1 km of the start of surface flow, and more than 700 km of tributaries are within concrete), have reduced sediment output from the river to 200,000 m^3 per year. This has severely exacerbated coastal erosion and loss of beaches (Flick, 1993). In this example, reduced river longitudinal connectivity affects coastal environments.
- More than 100 million tons of silt historically carried by the Nile River each year to the Mediterranean Sea nourished marine plankton communities at the river's mouth. Sardines fed on the plankton. Completion of the High Aswan Dam in 1970 dramatically decreased water and sediment yield from the Nile drainage basin to the Mediterranean, and contributed (along with pollution) to the collapse of the commercial sardine fishery (Collins, 2002). Diminished sediment supply has also resulted in subsidence and erosion of the Nile delta, where former delta villages are now 2 km out to sea (Stanley et al., 2004). Here, reduced river longitudinal connectivity alters delta and nearshore environments.
- The Novosibirsk Dam has altered flow along more than 1000 km of the upper and middle Ob River in Siberia since the dam's completion in 1957 (Bityukov, 1990). The Ob historically provided one of Russia's most important fisheries, partly because of the abundant floodplain habitat along the river. Annual overbank floods during snowmelt provided fish spawning habitat in floodplain meadows during May, and young fish could grow rapidly in the warm, slow-moving, shallow waters during June and early July. A minimum of 20 days of flooding each year are required for fish spawning, hatching, and growth along the Ob, but flow regulation associated with the Novosibirsk Dam reduces floodplain spawning and feeding areas by half in years of average flow, and completely during years of drought (Wohl, 2011a). Catches of commercially valuable species such as Siberian sturgeon (*Acipenser baerii*) dropped to nearly nothing by the mid-1990s (Ruban, 1997). In this example, reduced lateral connectivity alters riverine biota.
- North America's Mississippi River integrates land uses and human alterations of river process and form across 3.5 million km^2. Among the many substantial, human-induced changes in the drainage basin during the past century have been the loss of naturally vegetated floodplains and riparian zones, and the construction of hundreds of dams within the watershed. Loss of valley-bottom forests and wetlands and their

biogeochemical buffering effects, channelization and loss of the river-form complexity that promotes retention and biological processing of nutrients, and the spread of industrial agriculture and associated fertilizer and pesticide use during the latter half of the twentieth century, together created enormous increases in nutrients and contaminants transported from the river catchment into the Gulf of Mexico in the Atlantic Ocean (Goolsby et al., 1999). This has caused persistent hypoxia in the Gulf, and a "dead zone" that fluctuates between 13,000 and 20,000 km^2 in size (Mitsch et al., 2001). The Missouri River drains the western half of the Mississippi River basin, contributing relatively little water (12% of the Mississippi's discharge) but large amounts of sediment (>50% of the Mississippi's load) (Meade et al., 1990). Numerous large dams built along the Missouri since 1950 now trap about 80% of the sediment load. Along with other changes in land use, this reduced sediment discharge to the Gulf contributes to rapid delta and coastal erosion (69 km^2 of coastal wetlands lost each year, for an estimated total of 3900–5300 km^2 since the 1930s) (Turner, 1997). In this example, increased upland-river connectivity in the form of increased nutrient fluxes and decreased processing, and reduced longitudinal connectivity of sediment fluxes, substantially alter delta, coastal, and nearshore environments and biota.

- The Danube River drains 816,000 km^2 of Europe and, like the Nile and the Mississippi, changes in land use and flow regulation have negatively impacted delta and nearshore geomorphology, water quality, and biota. Compared to the 1960s, nitrogen loads entering the Black Sea from the Danube have increased fivefold, phosphorus loads have doubled, and silica has decreased by about two-thirds. The increased nitrogen and phosphorus reflect commercial agriculture and loss of delta lakes and floodplains that promote biogeochemical processing of nutrients. The result has been eutrophication of the lower river, delta, and nearshore areas. Decreased silica likely reflects the 70% drop in sediment yields to the delta as a result of widespread dam construction upstream. Declines in silica have caused Black Sea phytoplankton communities to shift from silica-using diatoms to coccolithophores and flagellates that do not require as much silica. This has caused algal blooms that destabilize the Black Sea ecosystem (Humborg et al., 1997). Eutrophication, algal blooms, and contaminants carried by river water have contributed to the collapse of once commercially valuable Black Sea sprat and anchovy fisheries. As in previous examples, increased lateral connectivity of nutrients and contaminants, and decreased longitudinal connectivity of sediment, have dramatically altered delta, nearshore, and coastal environments.
- The Fly River drains 75,000 km^2 in Papua New Guinea. The Ok Tedi, one of the Fly's major tributaries, originates in the steep, tectonically active Southern Fold Mountains, and includes the Ok Tedi mine. The open-pit copper–gold porphyry mine is the second largest copper-producing mine in the world. Production began in 1984. No tailings dam was constructed, and ~80,000 tons of waste tailings and 121,000 tons of mined waste rock are dumped directly into the Ok Tedi River each day, resulting in ~66 million tons of extra sediment per year (Hettler et al., 1997). Metals travel downstream in dissolved and particulate form (Yaru and Buckney, 2000). Elevated levels of copper have been detected in floodplain lakes (Nicholson et al., 1993) and in the Gulf of Papua beyond the mouth of the Fly River (Apte and Day, 1998). The bed of the Ok Tedi aggraded over 6 m in places during the decade following the start of mining (Swales et al., 1998). Mining sediment is also dispersed across the floodplain. Although direct overbank flooding and vertical accretion are limited to less than 1 km on either side of the main channel, sediment is dispersed across tens of kilometers of floodplain by backflooding of the mainstem up tributaries and secondary channels connecting the tributaries to floodplain wetlands (Dietrich et al., 1999, 2007). In this example, disposal of mining wastes artificially increased hillslope–channel connectivity, resulting in widespread downstream dispersal of sediment and metals, and altered river, floodplain, and nearshore environments.

The point of these examples of altered connectivity is not to induce despair as we contemplate past failures to account for the importance of connectivity, but rather to highlight the importance of being fully cognizant of various forms of connectivity as we move forward with river management. Rivers are physical systems with a history, and rivers exist within a global landscape that includes the atmosphere, land masses, oceans, groundwater, and all the wondrously diverse organisms that live on our planet. Rivers are at the heart of nearly every landscape on Earth, and river form and process are more integral to human communities than is any other single landscape component. The challenge of developing ways to co-exist with healthy, functional rivers is integral to both a scientific understanding of rivers and to applying that understanding through river management. This challenge can only be met by treating rivers as part of the greater landscape. If we do not view rivers in this integrative, holistic context, we make the same mistakes over and over and, ultimately, risk nothing less than the survival of our own societies.

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