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Andrew Miall

Fluvial Depositional Systems

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Fluvial Depositional Systems

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Andrew Miall
Department of Earth Sciences
University of Toronto
Toronto
ON
Canada

ISBN 978-3-319-00665-9 ISBN 978-3-319-00666-6 (eBook)
DOI 10.1007/978-3-319-00666-6
Springer Cham Heidelberg New York Dordrecht London

Library of Congress Control Number: 2013940939

Printed on acid-free paper

Springer is part of Springer Science+Business Media (www.springer.com)

Preface

In an earlier book, *The Geology of Fluvial Deposits* (1996), I set out in detail modern methods of facies and architectural analysis of fluvial deposits, and used numerous case studies to illustrate the architecture of fluvial systems, on scales ranging from that of the outcrop to that of entire basins. Chapters were devoted to the tectonic and climatic controls on fluvial deposition, and an attempt was made to erect a classification of nonmarine oil and gas fields based on the stratigraphy and architecture of the fluvial reservoirs.

In subsequent years, a host of new case studies has provided much material for refining our understanding of allogenic controls, and has substantially improved our ability to apply sequence-stratigraphic methods to fluvial systems. Exploration techniques used for petroleum exploration and development have become much more sophisticated, and in my view, are steadily reducing the need for much of the statistically based modeling work that is carried out during the reservoir development process, in favor of the detailed mapping of what is actually there, using such techniques as three-dimensional seismic reflection, and the careful analysis of production data, such as pressure-depth relationships.

One of the major foundations of sedimentological work has been the analogue method, whereby the processes and products of modern and very recent sedimentary environments form the basis for comparison with the ancient record. However, our increasing ability to develop accurate ages for the rock record has raised an important question about the validity of the analogue method, which forms the basis for one of the fundamental principles of geology, that of uniformitarianism. The fragmentary nature of preserved stratigraphies is increasingly apparent, and it is clear that comparisons to the ancient record based on studies of post-glacial stratigraphy, such as the great deltas and continental margin sedimentary prisms bordering modern oceans, must be carried out with a major caveat regarding questions of preservability. This is particularly the case in the area of sequence stratigraphy, an area that is examined in depth in this book as it relates to the analysis and interpretation of fluvial deposits.

The purpose of this book is to discuss the new methods and the new understanding of fluvial depositional systems, with a particular emphasis on those techniques and results that are most useful for subsurface work.

Toronto, March 2013 Andrew Miall

Acknowledgments

The background to this book has been formed by the numerous presentations of courses on fluvial depositional systems to petroleum geology audiences in Calgary and elsewhere. I have repeatedly asked myself the question, what do these professional people need to hear from me? I am grateful, particularly to the Canadian Society of Petroleum Geologists, for the opportunity to be forced to ask myself this question, and for the expectation that I can respond with relevance.

I am particularly grateful to Robin Bailey and David Smith for their invitation to speak at the *Geology and Time* symposium at the Geological Society of London in September 2012. The preparation for this occasion led me to explore an area that has long troubled me, that of sedimentation rates and stratigraphic completeness. The ideas contained in the chapter on sequence stratigraphy in this book owe much to the rethinking that went into my contribution to that event.

Conversations with many colleagues over the years have helped me to formulate and clarify my ideas, and now being able to access the research literature from even the most remote holiday cabin with Internet access means that, to some extent, the thinking never stops. In recent years I have found the work of Phil Allen, Robin Bailey, Janok Bhattacharya, Nick Eyles, Chris Fielding, Martin Gibling, John Holbrook, Colin North, Chris Paola, Guy Plint and Pete Sadler, particularly illuminating. Paul Heller critically read most of the manuscript and provided many essential comments.

As with all my work, conversations in the field and in the office with my graduate students have added immeasurably to my understanding of geological problems and my abilities to explain them. With regard to fluvial processes and systems I must mention, in particular (in alphabetical order), Tosin Akinpelu, Mike Bromley, Gerald Bryant, Octavian Catuneanu, Jun Cowan, David Eberth, Carolyn Eyles, Phil Fralick, Greg Nadon, Tobi Payenberg, Mark Stephens, Andrew Willis, and Shuji Yoshida.

Finally, I must again thank my wife Charlene for putting up with yet another book. This fluvial enterprise owes much to Charlene's field assistance, companionship, professional advice, support, and love over the years, and the work we did together on the history and methodology of stratigraphy owes much to her educating me about the nature of the scientific method and the sociology of science.

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Chapter 1
The Nature and Purpose of This Book

1.1 Looking Back to 1996

This book is not a revised version of "*The geology of fluvial deposits*" (Miall 1996), but an entirely different product.

Much of the material in the 1996 book was compiled at a time when the methods of facies analysis and architectural element analysis were maturing and were becoming widely used by the sedimentological community. The lithofacies classification which I first proposed in 1977, and the method of architectural-element analysis, set out in major papers published in 1985 and 1988, were thoroughly documented in the 1996 book (Chaps. 5–7), and little has been done since then to require revisions or an upgrade. A recent summation of the methods was provided by Miall (2010a). As expected, indeed, as was recommended, researchers have taken the basic ideas and adapted them to suit the particular needs of their research projects. Field techniques now include the use of LIDAR for the recording of outcrop images, which may substitute for photomosaics, but the methods of outcrop architectural analysis (Chap. 4) remain much the same. New approaches and techniques for mapping the subsurface have been developed for use in the petroleum industry; these are introduced briefly below and discussed at greater length in Chap. 4 of the present book. Three-dimensional reflection-seismic data is increasingly becoming a standard tool for petroleum geologists, and its interpretive arm, seismic geomorphology, is a powerful tool requiring a deep knowledge of sedimentology for maximum usefulness.

The compilation of facies models that constituted Chap. 8 of the 1996 book has largely stood the test of time. Only one new facies model has been formally proposed since that time, a model for rivers in hot, seasonal, semiarid and subhumid environments (Fielding et al. 2009, 2011). Extensive research, such as that by Long (2011) has demonstrated the applicability of the original suite of models to the rock record.

A. Miall, *Fluvial Depositional Systems*, Springer Geology,
DOI: 10.1007/978-3-319-00666-6_1,

The tectonic control of fluvial systems was thoroughly described in Chap. 11 of the 1996 book, and the chapters dealing with oil and gas fields in fluvial systems (Chaps. 14 and 15) need little modification.

For a research-level textbook covering all this material, the reader is still referred to the 1996 book.

1.2 New Developments

The area that appeared to require the most extensive revision and renewal is, not surprisingly, the material dealing with sequence stratigraphy (Chap. 13 in the 1996 book). Much has changed since that chapter was written, and indeed, whole new ways of thinking have evolved that require some new approaches. Some of these new ways have thinking have developed from the imaginative and quite revolutionary laboratory work undertaken by Chris Paola at his experimental facility at the University of Minnesota. In this research, fluvial and deltaic processes have been modeled in a large tank that has been constructed to simulate base-level change and differential subsidence. Theoretical arguments and comparisons with modern fluvial-deltaic systems have established that the results of the experiments may be scaled up to that of natural systems, thereby filling an essential observational gap, termed the "mesoscale", between the documentation of modern and historical processes, which essentially only cover about the last 100 years, and geological observations on the rock record, for which the most refined time scale available is that of magnetostratigraphy, in the 10^4-year range. Results and conclusions drawn from the work of Paola and his group have been integrated into the discussion at several places throughout this book.

Another critical development in the last two decades has been the steady accumulation of quantitative data relating to sedimentary and stratigraphic processes. We now know substantially more than we did in the 1990s about the rates of sedimentary processes, and about the nature and rates of an increasingly wide range of allogenic forcing processes. Work by such researchers as Paul Heller, Doug Burbank, David Mohrig and Elizabeth Hajek, amongst others, has aimed to test experimental and theoretical work against observations from the ancient fluvial record, using carefully selected field case studies. Some of these results are discussed in this book, focusing primarily on the larger-scale fluvial systems and those components of which (channels, channel belts and depositional systems) that are the main focus of the subsurface geologist.

However, in one important area, this increasingly detailed and quantitative knowledge of sedimentary processes has led to what, in this writer's view, is the emergence of a serious but hitherto largely unrecognized disconnect between those studying modern processes and the post-glacial record, versus those studying the more ancient record. The increasingly large data base that is now available to researchers on rates and time scales has demonstrated that sedimentation rates, and the rates of processes, as measured in modern environments and in

Pleistocene-Holocene settings, are one to three orders of magnitude more rapid than those normally derived from detailed chronostratigraphic studies of the pre-Neogene record. This is partly a reflection of the high rates and large magnitudes of changes that occurred through the late Cenozoic glacial cycles, but it is also a reflection of the nature of what I have called the "geological preservation machine", whereby high-frequency, high-rate events are systematically removed from the record as time passes. This is not a new observation; it was an obvious conclusion of the work of Sadler (1981) who published a by-now classic paper on sedimentation rates (Fig. 1.1). What is new is the availability of new theory and much new data to assist in the explanation of this phenomenon. As discussed in Chap. 2, and at greater length elsewhere (Miall, in press), stratigraphic processes over the full spectrum of geological time scales may now be understood with reference to the concept of fractals.

In a widely-quoted remark, the implications of which have been largely ignored in practice, Ager (1973) stated that "the stratigraphic record is more gap than record." As Miall (in press) argued, we now have to consider the fact that there are, in effect, gaps within the gaps, and that the record is permeated with them,

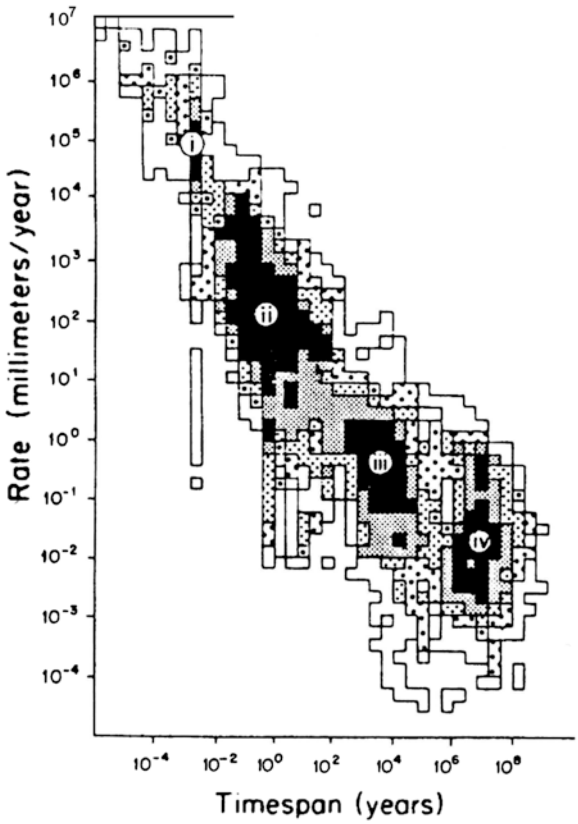

Fig. 1.1 The relationship between sedimentation rate and elapsed time in the stratigraphic record (Sadler 1981)

at every scale. Preserved stratigraphy constitutes a set of fragmentary remnants, that have been called "frozen accidents." (Bailey and Smith 2010, pp. 57–58). These can tell us a great deal, but only if we work within the appropriate time scale. Much of the present book consists of a working through of the implications of these concepts for fluvial systems and the fluvial sedimentary record. This constitutes part of what I have called "*updating uniformitarianism*" (Miall, in press).

We need a new approach to uniformitarianism, because of the disconnect, noted above, between those working on the modern and the ancient record. It could be argued that the analog method on which modern sedimentology is based, is no longer a satisfactory foundation for research into long-term geological processes. It was based on the long-standing, traditional Hutton-Lyell aphorism "the present is the key to the past", and its obverse, "the past is the key to the present". If the geological preservation machine systematically removes much of the modern record before it can become part of the ancient record, we need to be constantly alive to the potential for the bias this introduces into our interpretations.

In the practical world of petroleum geology, the transition that takes place from exploration to production involves a handover from the geologist to the engineer of a model of reservoir architecture to be used as the basis for the design of a production program. There are tensions in this process because of the level of uncertainty inherent in geological prediction (e.g., Martin 1993). Speaking of high-risk and high cost frontier exploration project, Larue and Hovadik (2008, p. 337) said:

> Project appraisal and development may be based on very few wells with or without the benefit of 3D seismic data, but with implications for capital costs of hundreds of millions to billions of dollars.

The qualitative nature of these models may not satisfy the quantitative requirements of the engineer. Typically this is now managed by the use of numerical models that employ probabilistic methods to provide ranges of likely values for engineering purposes. Essential information, such as the dimensions and spacing of reservoir bodies may be calculated as ranges of likely values from the sedimentological and sequence models compiled by the geologist. There are many commercial computer modeling methods that manage this part of the production process. With the exception of the next, concluding paragraph of this section they are not discussed in this book, the main purpose of which is to assist the geologist to understand the fluvial system from which the computer input is assembled.

A specialized area of computer simulation has grown to answer the following problem:

> In industry scenarios, the typical paucity of data relating to sedimentary heterogeneity at a resolution finer than the seismic and interwell-spacing scales, together with the need to undertake uncertainty analysis for the assessment of risk, has resulted in the need for the development and implementation of stochastic methods for modeling reservoir sedimentary architecture by simulating several different equiprobable architectural realizations. Structure-imitating stochastic reservoir modeling aims at simulating sedimentary architecture without considering depositional and/or erosional processes.

This quote, from Colombera et al. (2012, p. 2144) introduces an elaborate new database system from which to sample input parameters relating to depositional systems, architectural elements and lithofacies in order to construct reservoir models for development engineering purposes. This approach appears to be by far the most sophisticated in this category of model building. The purpose is not to simulate fluvial processes, but to construct a practical architectural model for reservoir planning purposes based on the limited input data available from preliminary exploration and interpretation of facies, fluvial style, tectonic and climatic setting, etc. However, as discussed throughout this book, fluvial systems are notoriously difficult to predict. The simple case of trunk rivers having tributaries of variable scales and fluvial styles (Fig. 2.11) may play havoc with a well-thought-out engineering model. It is to be hoped that the contents of this book can assist in the work to understand and constrain the input that needs to be used in models of this type or, as we discuss in Chap. 4 (see below) to try to avoid statistical approaches altogether, as much as possible, by the employment of various new mapping tools.

1.3 Introduction to the Contents of This Book

Chapter 2: Modern fluvial sedimentology began with the development of the point bar model and the fining-upward cycle in the late 1950s and early 1960s (Miall 1996, Chap. 2). The process-response model flourished in the subsequent decades, and has left us with a wealth of information on modern rivers and ancient deposits, much of it categorized under the heading of facies models. In Chap. 2 I take a look at the modern state of fluvial facies studies, and conclude that the facies model approach has long-since reached its limit of usefulness. One of the difficulties is the selective preservation of modern fluvial processes. For example, studies of the shallow deposits of modern rivers using ground-penetrating radar have demonstrated that the surface form is often not reflected by the internal structure, but is superimposed on fragments of earlier channel and bar deposits above recent local erosion surfaces. This is part of the preservability issue that I raised earlier. Another important point is the growing data base that points to the low level of predictability that can be inferred from geological studies of the rock record. Gibling's (2006) survey of dimensional data on fluvial facies units is examined in Chap. 2, where I reproduce some of his data documenting such relationships as the width:depth ratio. These kinds of relationships have been used for a long time as predictive tools for studying the subsurface, but are not, in fact, very discriminatory. Prediction of subsurface dimensions from limited vertical profile, including core, data, is fraught with hazard.

Architectural methods of description and documentation of the rock record are more powerful than traditional vertical-profile methods, because they direct the observer to seek out three-dimensional information, and come with no presumptions regarding fluvial style. However, the methods are difficult if not impossible to use where there is only limited well data, as in the early phases of a subsurface exploration program. In addition, many of the most interesting architectural elements, such as nested channels and incised valleys, are commonly at scales of hundreds of metres to a few kilometres across that commonly are too large to be seen completely in outcrop or sampled reliably by exploration wells, but too small to be seen properly on reflection seismic data. I discuss this problem further in Chap. 6.

Chapter 3: A major concern of the subsurface geologist is the problem of defining and describing the architecture of the various facies, particularly the porous units—typically composed of sandstone or conglomerate, that constitute potential or actual petroleum reservoirs. The size, orientation and connectivity of these bodies are critical to the effective and efficient design of well networks, particularly for the purpose of enhanced recovery projects. Much depends on the ways in which fluvial channels move around on a floodplain, whether by gradual migration or by sudden shifting—the process termed *avulsion*—the major focus of this chapter. Geological work on this problem has consisted of extensive study of the history of avulsion of modern rivers, mapping of ancient avulsions in the rock record, and the numerical and experimental modeling of avulsions. The physical processes of avulsion are complex, and are still not completely understood. Numerical models of the avulsion process, of which there are several, do not attempt to simulate the physics of the process and are essentially exercises in dynamic geometry. The results of laboratory experiments, primarily those of Chris Paola's *experimental stratigraphy* laboratory, are helping to throw light on the issue. Despite decades of activity in this area, a definitive treatment of the issue of avulsion, and the more general topic of the autogenic control of alluvial architecture, is still not possible.

Of key practical importance to the business of reservoir development is the nature of the sand fairway. Sand body connectivity is the key descriptor, and in this chapter we discuss the critical factors on which it depends. It can be demonstrated that fluvial style is NOT a critical element in the determination of reservoir performance.

Chapter 4: Moving on to larger-scale features of fluvial systems, where allogenic processes become predominant, requires the construction of detailed maps and sections of fluvial systems. Modern mapping methods (Table 1.1) include a range of dynamic tools that make use of production measurements, and are more effective than the traditional methods based on the facies model and the vertical profile, in that they are empirical, directed towards systematically revealing what is actually there rather than attempting to predict based on assumed relationships that may have little factual, basis. Some of these methods make use of the dynamic production data that may be collected as a petroleum field is developed.

Chapter 5: Developments in the understanding of tectonic and climatic control of fluvial sedimentation have advanced significantly in the last few decades owing to the accumulation of numerous case studies. The improved understanding of high-frequency tectonism in foreland basins, and a much broader knowledge of the development of paleosoils, including their dependence on climatic controls, are two developments that have significantly improved our range of tools for interpreting the ancient fluvial record. At the same time, experimental and theoretical research have provided essential insight into rates and scales, particularly regarding such issues as the response time of alluvial systems to allogenic forcing.

Table 1.1 Methods for mapping complex fluvial systems in the subsurface

Old/traditional methods largely based on facies-models concepts	
"*The geostatistics of random sandstone encounters*"	*Discussed in Sects.:*
The vertical profile (and its limitations)	2.2.1–2.2.2
Width-depth ratios and other geomorphic relationships	2.2.3
Architectural elements	2.3
Idealized bar models	
Net: gross and sandbody connectivity	3.6
Reservoir models and their limitations	
Newer, empirical methods:	
Ground-Penetrating Radar (GPR)	4.1.2
3-D seismic surveys	4.2.1
Dipmeter and formation microscanner	Miall (1966, Sect. 9.5.8)
Dynamic methods:	
"*Stroking the substrate*" *with directional drilling*	
Pressure testing	4.2.2
Geochemical fingerprinting, tracer testing, etc.	4.2.2
4-D seismic surveys	4.2.2
History matching	4.2.2

Chapter 6: Two important sequence models that were developed for fluvial deposits in the 1990s have been very influential, but in Chap. 6, I suggest that they commonly have been misapplied to the rock record. A number of worked examples are used to illustrate the argument that because these models are largely based on observations from modern rivers and the post-glacial sedimentary record, they cannot be applied directly to the ancient record, because of the issues of sedimentation rates and preservation, that I introduced above.

There has been much interest in fluvial sequence boundaries in the last few years, particularly the way in which erosional boundaries develop through lengthy periods of negative accommodation. The shaping of this surface and the fragmentary deposits that are commonly left behind during this process provide a graphic insight into the succession of vanished landscapes that evolve during these periods—and suggest an illustration based on modern data of the "abyss of time" that was so eloquently described by John Playfair on seeing Hutton's angular Silurian-Devonian unconformity at Siccar Point for the first time. As with other aspects of fluvial processes, the experimental stratigraphy experiments of Paola's group are providing many useful insights.

Lastly, in Chap. 7, I discuss the issue of identifying large rivers and their associated depositional systems in the rock record. There has been a substantial recent literature published on the matter of large rivers (Gupta 2007; Ashworth and Lewin 2012), large-scale depositional systems (Weissmann et al. 2010, 2011; Fielding et al. 2012), and paleovalleys (Gibling et al. 2011; Blum et al. 2013). Much of this is focused on rivers and valleys of the present day and the post-glacial period, but there are limits on how far these data can be applied to the task of reconstructing ancient depositional systems. Modern and recent systems can be interpreted in terms of contemporaneous tectonism and climate, but when studying the ancient record, the problem is the reverse: that of deriving the maximum amount of information from what is often very fragmentary and incomplete evidence—evidence that is commonly quite ambiguous. One of the outstanding issues dealt with, in particular, by Gibling et al. (2011), is the problem of discriminating between paleovalleys and large channel systems.

1.4 Conclusions

The main purpose of this book is to assist those working with the rock record to maximize the information they can obtain from their research. Architectural methods have contributed substantially to the interpretation of preserved fluvial systems at the outcrop scale. In the case of the subsurface—the attempt to map and explain potential reservoir units or to provide more complete descriptions of producing units—many of the same problems remain as they have been for decades: the limitations on interpretation that are imposed by the lack of critical data. However, where available, such new exploration tools as the 3-D seismic-reflection method, and some mapping methods that make use of production data, can add substantially to the depth and reliability of interpretations.

As background to all of this are developments in our understanding of the "geological preservation machine", the means by which allogenic and autogenic processes operate over an enormous range of time scales to create the preserved rock record, with all its recognizable features, such as channel systems and sequences, while also inserting subtle and not so subtle gaps in the record, that make the work of the geologist continually challenging.

Chapter 2
The Facies and Architecture of Fluvial Systems

2.1 Introduction

In Sect. 2.3.1 I pose the question: why do petroleum geologists worry about fluvial style? and provide the answer: it is because it has long been assumed that reservoir architecture is the key to reservoir performance. In this chapter we discuss some of the difficulties in the reconstruction of fluvial style and facies architecture from the ancient rock record. It is important to note, however, that reservoir architecture, as such, may not be the critical key to reservoir performance that it has commonly been thought to be. As Larue and Hovadik (2008) have demonstrated, from their series of numerical experiments, facies variation along the flow paths, and its control on permeability, is of the greatest practical importance. The most important control on reservoir performance is sand body connectivity (the "sand fairway"), which may only be loosely dependent on reservoir architecture. Channel density and stacking pattern, regardless of the style of the channels, are the key controls on connectivity. Sand body connectivity is discussed in Sect. 3.7.

2.2 Depositional Scales

One of the most distinctive features of the earth sciences is the wide range of scales with which we have to deal (Fig. 2.1). The concept of deep time is a concern of earth scientists, theoretical physicists and astronomers. On Earth we deal with 4.5 billion years of time (about one third of the duration of the universe), but we deal with it in different ways on different time scales that vary over sixteen orders of magnitude:

- The formation of continents, basins and basin-fill successions over millions to as much as a billion years;
- The effects of tectonism and climate change on time scales of 10^4–10^7 years;

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DOI: 10.1007/978-3-319-00666-6_2,

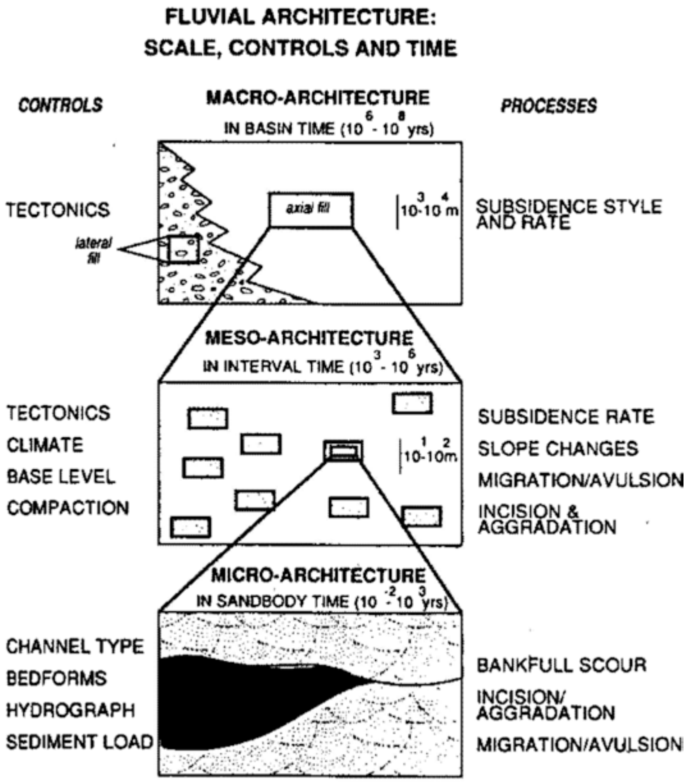


Fig. 2.1 Hierarchies of scale and time in fluvial deposits (Leeder 1993)

– The evolution of depositional systems, a geomorphic process that addresses processes over a time scale of tens to hundreds of thousands of years;
– The formation of bedforms and local aggradational cycles in response to daily and seasonal processes and to dynamic events (e.g., the 100-year flood). These processes are observable in present-day depositional systems, but for the purpose of understanding the ancient record we need to be aware that most of what we observe is geologically ephemeral.

It has become a geological truism that many sedimentary units accumulate as a result of short intervals of rapid sedimentation separated by long intervals of time when little or no sediment is deposited (Ager 1981, 1993). It is also now widely realized that rates of sedimentation measured in modern depositional environments or the ancient record vary in proportion to the time scale over which they are measured. Sadler (1981) documented this in detail, and showed that measured sedimentation rates vary by eleven orders of magnitude, from 10^{-4} to 10^{7} m/ka (Fig. 1.1). This wide variation reflects the increasing number and length of intervals of nondeposition or erosion factored into the measurements as the length of the measured stratigraphic record increases. Breaks in the record include such events as the nondeposition or erosion that takes place in front of an advancing bedform (a few seconds to minutes), the nondeposition due to drying out at ebb tide (a few hours), up to the major regional unconformity generated by orogeny (millions of years).

The variation in sedimentation rate also reflects the variation in actual rates of continuous accumulation (fifteen orders of magnitude in total), from the rapid sandflow or grainfall accumulation of a cross-bed foreset lamina (time measured in seconds, or 10^{-6} years), and the dumping of graded beds from a turbidity current (time measured in hours to days), to the slow pelagic fill of an oceanic abyssal plain (undisturbed in places for hundreds or thousands of years, or more), to the development of a major structural-stratigraphic province, which could represent hundreds of millions of years. There clearly exists a wide variety of time scales of sedimentary processes (Figs. 2.1, 2.2, 2.3).

There also exists a hierarchy of physical scales, which the same two examples illustrate—the cross-bed foreset at one extreme to the basin-fill at the other extreme (Fig. 2.1). At least fifteen orders of magnitude are represented, from the few square centimeters in area of the smallest scale of ripple foreset, to the tens of thousands of square kilometers of a major sedimentary basin. At the scale of the bedform, physical scales are constant, because they reflect invariant processes of

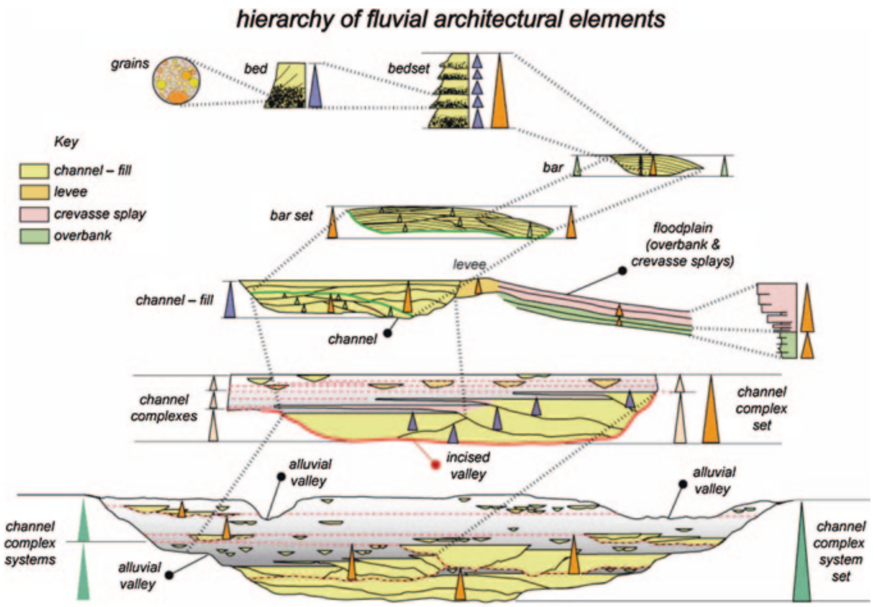


Fig. 2.2 The hierarchy of depositional units in a fluvial complex. This diagram was developed primarily to assist in the explanation of sequence-stratigraphic terms and concepts (Kendall 2008; sepmstrata.org)

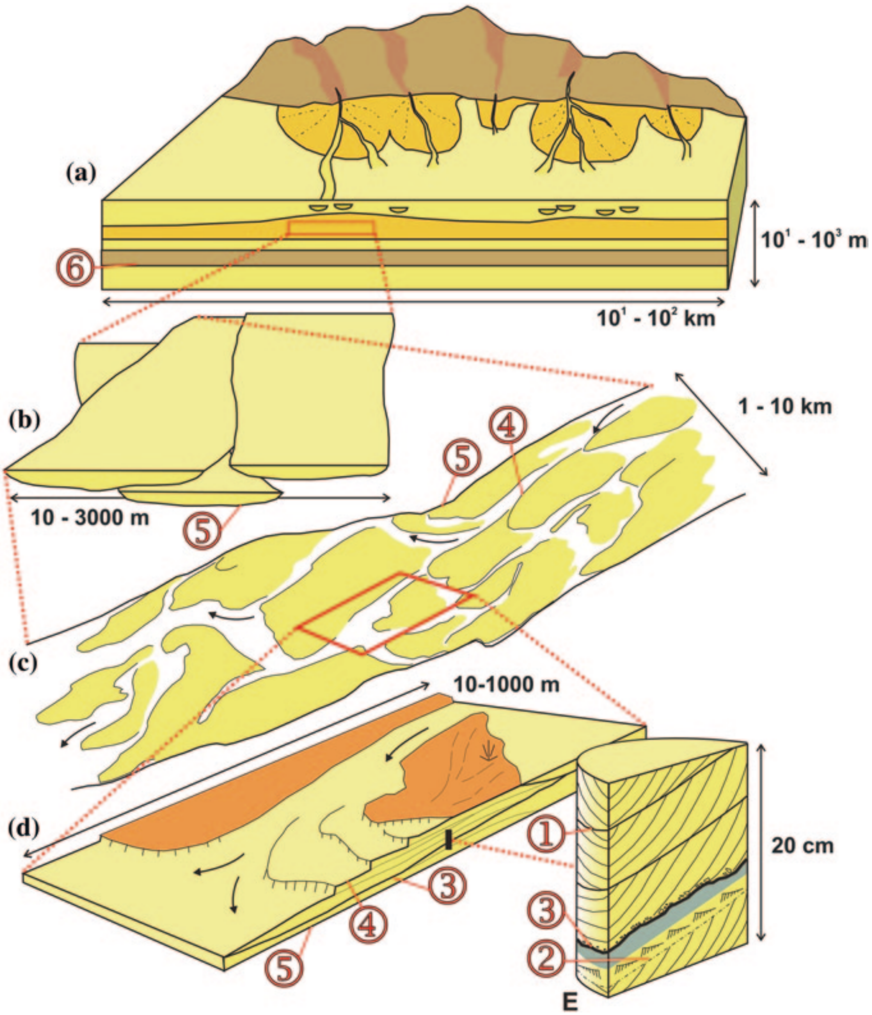


Fig. 2.3 The hierarchy of depositional units in a fluvial system. *Circled numbers* indicate the ranks of bounding surfaces, using the classification of Miall (1996)

the physics of sedimentation. However, at other levels of the hierarchy the scales may show wide variation, such as the scales of fluvial channels (Fig. 2.4).

The ways by which earth scientists study sedimentary processes and the resultant depositional products vary according to the scale of interest (Table 2.1). Bedforms in flumes are studied during experimental runs of, at most, few days duration. Nonmarine and marginal-marine sediments and processes have been much analyzed in modern environments, using studies of surface processes, and by sampling the sediments themselves in trenches and shallow cores. The use of old maps and aerial photographs extends the record as far back as about 100 years.

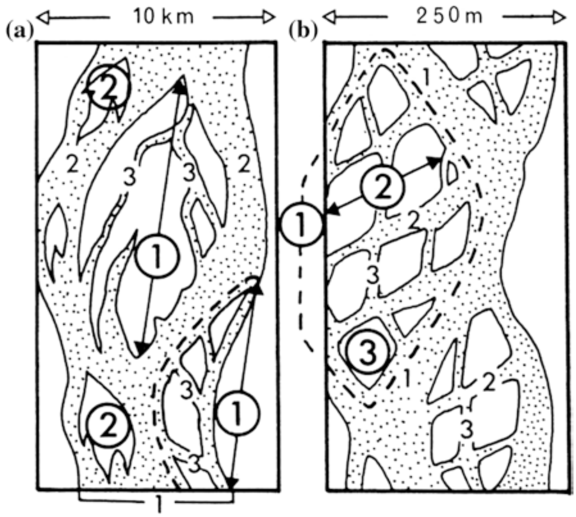


Fig. 2.4 Channel hierarchies in the Brahmaputra River, (**a**), and the Donjek River, (**b**) (after Williams and Rust 1969). *Numbers in circles* refer to bars, *other numbers* refer to channels. The first-order channel comprises the whole river, which includes several second-order channels. Bars scale within the channels in which they occur. In the Brahmaputra River third-order channels modify higher-order bars but still have bars within them, which cannot be shown at this scale (Bristow 1987)

Optically stimulated luminescence (OSL) can provide age information for the 300-100,000-BP time span. ^{14}C dates may enable stratigraphic records of the last few tens of thousands of years to be calibrated. Many sedimentological studies draw on geomorphological work on landforms and Recent sediments. However, such work is hampered by the specific, and possibly non-generalizable nature of the Recent record, such as the Holocene deglaciation, climatic change, and rapid rise of global sea levels. Stratigraphic studies typically deal with much longer time periods, as represented by the deposits of basin fills, which may have taken hundreds of thousands to millions of years to accumulate. Intermediate scales, represented by such major depositional elements as large channels and bars, delta lobes, draas, coastal barriers and shelf sand ridges, which may represent thousands to tens of thousands of years of accumulation, are particularly difficult to document in the ancient record and to analyze in modern environments. The time scales of the relevant sedimentary processes are difficult to resolve, and the physical scale of the deposits falls between the normal size of large outcrops and the well spacing or the scale of geophysical resolution in the subsurface. Yet it is this scale of deposit that is of particular interest to economic geologists, representing as it does the scale of many stratigraphic petroleum reservoirs and their internal heterogeneities (Table 2.2).

Figures 2.2 and 2.3 represent two different ways of illustrating the hierarchical nature of stratigraphic accumulations. Most of the problems faced by geologists attempting to wrestle with field-scale heterogeneities relate to the intermediate scales shown on these diagrams, the channel fill and channel complex of Fig. 2.2, and the units shown in diagrams B, C, and D of Fig. 2.3. We return to these scale issues in a discussion of reservoir problems, in the next section.

Geomorphologists have devoted considerable attention to the problem of time scales and their effects on analysis and prediction (Cullingford et al. 1980;

Table 2.1 Hierarchies of architectural units in fluvial deposits

SRS	Time scale	Inst. sed. rate	Examples of processes	Depositional unit	Type of process	Interpretive significance	Investigative technique
1	10^{-6}	10^{6}	Burst-sweep cycle	Lamination	Autogenic	Trivial	Thin-section hand specimen
2	10^{-6} –10^{4}	10^{5}	Ephemeral flow events	Ripple (microform)	Autogenic	Superficial hydraulic fluctuations	Hand specimen core
3	10^{-3}	10^{5}	Diurnal dune increment, reactivation surface	Diurnal	Autogenic	Daily variability small outcrop	Core
4	10^{-2} –10^{1}	10^{4}	Storms (mesoform)	Dune	Autogenic	Dynamic events	Core, small outcrop
5	10^{0}–10^{1}	10^{2}–10^{3}	Seasonal to 10 year flood	Macroform growth increment	Autogenic	Major dynamic events	Large outcrop GPR on modern river
6	10^{2}–10^{3}	10^{2}–10^{3}	100 year flood levee, splay	Macroform, e.g. point bar	Autogenic	Major dynamic events	Large outcrop GPR on modern river
7	10^{3}–10^{4}	10^{0}–10^{1}	Avulsion	Channel	Autogenic	Behavior of river system	Large outcrop horizontal 3-D seismic section
8	10^{4}–10^{5}	10^{-1}	5th order (Milankovitch) cycles	Channel belt	Autogenic or allogenic	Geomorphic response to regional change	Regional outcrop network horizontal 3-D seismic section
9	10^{5} –10^{6}	10^{-2} –10^{-1}	4th order (Milankovitch) cycles	Depositional system, alluvial fan, major delta	Allogenic	Tectonism, climate change, base-level change	Regional 2-D seismic or well network
10	10^{6}–10^{7}	10^{-1} –10^{0}	2nd–3rd order cycles	Basin-fill complex	Allogenic	Rapid tectonism	Regional 2-D seismic or well network
11	10^{6}–10^{7}	10^{-2}–10^{-1}	2nd–3rd order cycles	Basin-fill complex	Allogenic	Tectonism	Regional 2-D seismic or well network
12	10^{6}–10^{7}	10^{-3}–10^{-2}	2nd–3rd order cycles	Basin-fill complex	Allogenic	v. slow cratonic subsidence	Regional 2-D seismic or well network

Adapted from Miall (1996, in press), with ideas from Brierley (1996). GPR = ground penetrating radar

Table 2.2 Classification of fluvial-channel bodies and fluvial-valley fills according to size and form (Gibling 2006)

Width (m)		Thickness (m)		Width/Thickness		Area (km^2)	
Very wide >	10,000	Very thick >	50	Very broad Sheets >	1,000	Very large >	10,000
Wide >	1,000	Thick >	15	Broad sheets >	100	Large >	1,000
Medium >	100	Thick >	15	Narrow sheets >	15	Medium >	100
Narrow >	10	Thin >	1	Broad ribbons >	5	Small >	10
Very narrow <	10	Very thin <	1	Narrow ribbons <	5	Very small <	10

Hickin 1983; Schumm 1985a). As Hickin (1983, p. 61) has stated, "time-scale selection largely determines the questions that we can ask." Schumm (1985a) showed that the significance of an event diminishes as the time-scale increases. Thus, an individual volcanic eruption, a spectacular geological event at the time of its occurence (a "megaevent", to use Schumm's term), diminishes in geological importance as the millenia go by and other eruptions take place, until eventually, after perhaps millions of years, all evidence of the eruption is lost (it becomes a "nonevent") as a result of erosion or burial of the rocks and landforms formed by the eruption. Events that seem random in the short term (such as turbidity-current events) may assume a regular episodicity, or even cylicity, with definable recurrence intervals, if studied over a long enough time scale. Many events occur only when some critical threshold has been passed, such as the buildup of deposits on a depositional slope leading to gravitational instability and failure. In several essays, Schumm (1977, 1979, 1985a, 1988; Schumm and Brakenridge, 1987) has discussed the concept of "geomorphic thresholds" and their impact on sedimentary processes. Such thresholds reflect both autogenic and allogenic processes, and are characterized by a wide range of time scales (Fig. 2.5) and scales of cyclicity (Fig. 2.6).

The accumulation of information relating to sedimentation rates and its interpretation based on fractal theory has led to two important developments: (1) The realization that the stratigraphic record is far more fragmentary than has hitherto been appreciated (Miall, in press); and (2) The realization that many processes are scale independent. This has been argued from the perspective of sequence stratigraphy (Posamentier et al. 1992; Catuneanu 2006). It has also been argued from the basis of experimental and theoretical considerations that small-scale experiments, such as those carried out in the Experimental Earthscape Facility (XES) at University of Minnesota can be used to explore full-scale sedimentological processes that take place over geologically significant periods of time (Paola et al. 2009).

Miall (in press) proposed the definition of a suite of *Sedimentation Rate Scales* to encompass the range of time scales and processes that can now be recognized from modern studies of the stratigraphic record (Table 2.1, Fig. 2.7). Assignment of stratigraphic units to the appropriate scale should help to initiate a potentially rich new form of debate in which tectonic and geomorphic setting, sedimentary processes and preservation mechanisms can be evaluated against each other,

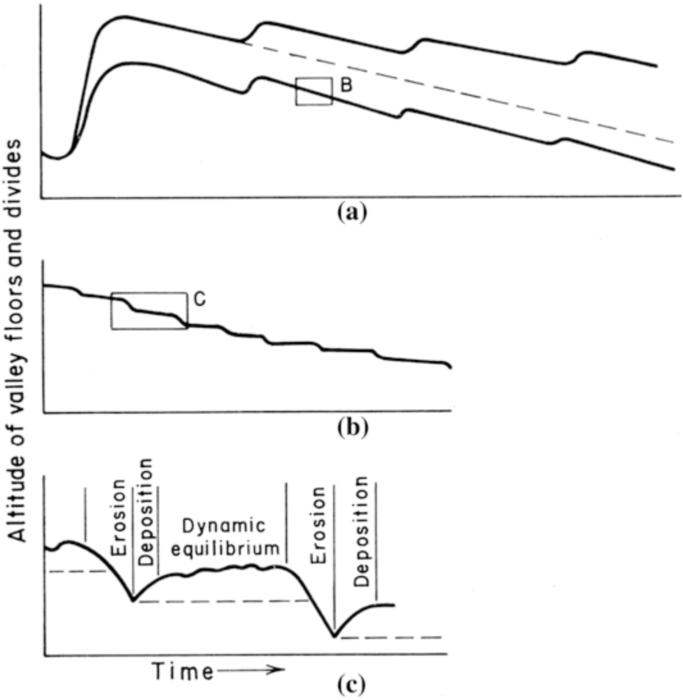


Fig. 2.5 The various time scales of geomorphic processes. **a** The erosion cycle, as envisioned by W. M. Davis in the nineteenth century. The *lower line* indicates the elevation of the valley floor, the *upper line* that of drainage divides. Initial uplift is followed by degradational lowering and episodic pulses of isostatic uplift in response to erosional unroofing. Total elapsed time is in the order of 10^{7-8} years for a major drainage basin, with minor uplift events occurring on the scale of 10^{6-7} years (corresponding to the tectonic cyclothems of Blair and Bilodeau (1988)). Box labelled B is enlarged in diagram (**b**). In detail the valley floor shows an episodicity on a smaller time scale (in the range of 10^{2-3} years) as a result of the periodic storage and flushing of sediment from bars and floodplain deposits, for example by avulsion events. Box labelled C is shown enlarged in (**c**), in which the episodicity of diagram (**b**) is shown in greater detail (diagram from Schumm 1977)

leading to more complete quantitative understanding of the geological preservation machine, and a more grounded approach than earlier treatments of "stratigraphic completeness".

The incorporation of hierarchical scale concepts into fluvial studies requires an architectural approach. Early approaches to the architectural study of fluvial deposits, notably, the work of J. R. L. Allen and of A. Ramos and his colleagues, is described elsewhere (Miall 1996, Chap. 2). The main classification used in this book is briefly described in Sect. 2.3. The current explosion of interest in sequence stratigraphy represents an increasing interest in large-scale stratigraphic architecture, and its dependence on such allogenic controls as tectonics and sea-level change, which topics form one of the major focuses of the present book (Chap. 6).

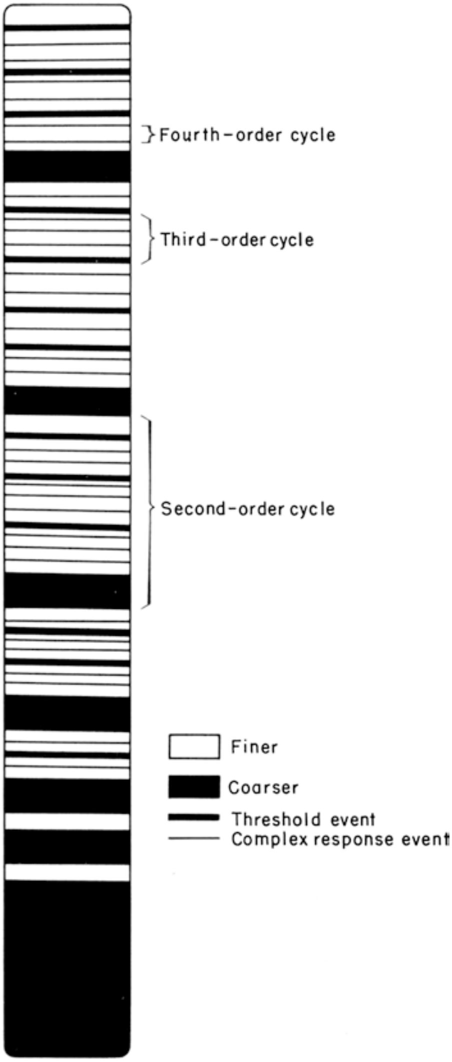


Fig. 2.6 The hierarchy of cycles of sedimentation, based on geomorphic concepts of the complex and episodic response of fluvial systems to autogenic and allogenic forcing. Schumm's cycle terminology does not correspond to that which emerged with sequences stratigraphy (Vail et al. 1977), and is explained here with reference to the *Sedimentation Rate Scales* of Table 2.1. The primary cycle is the entire succession, reflecting the gradual diminution of sediment grade following initial uplift (corresponding to the "erosion cycle" curve of Fig. 2.5a; *SRS 10* of Table 2.1). Second-order geomorphic cycles reflect isostatic adjustments (tectonic cyclothems) or major climate change (the kinks in the curves of Fig. 2.5a; corresponding to *SRS8-10* of Table 2.1). Third-order geomorphic cycles are those relating to the exceeding of geomorphic thresholds, leading to periods of "metastable equilibrium" and periods of rapid change and adjustment (The events shown in Fig. 2.5b). These processes occur over various time scales (groups 6–8). Fourth-order cycles are related to episodic erosion, and to the complex response of the fluvial system to any of the above changes (*SRS 5–8*). Fifth-order cycles are related to seasonal and other major hydrological events, such as the "hundred-year flood" (*SRS 5, 6*) (Schumm 1977)

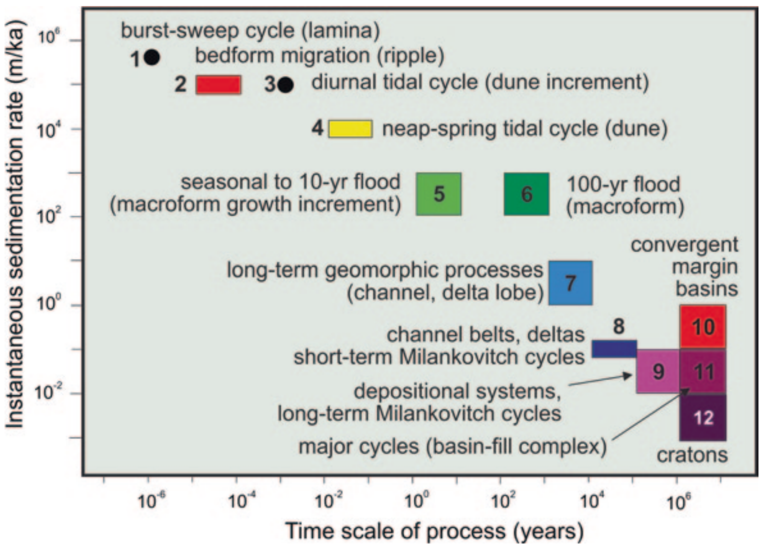


Fig. 2.7 Rates and durations of sedimentary processes. Numerals refer to the *Sedimentation Rate Scale* (see also Table 2.1)

2.3 Fluvial Style

2.3.1 Statement of the Problem

A great deal of sweat and much ink has been spent on worrying about fluvial style, that is, the shape and arrangement of channels on the valley floor of a fluvial system. Why? Because it has long been thought that fluvial style is the key to reservoir architecture. Until the advent of three-dimensional seismic, and the emergence of seismic geomorphology as practical tools for exploration and development of stratigraphic traps, geologists had very little data and only very unreliable tools to reconstruct reservoir geometry in the subsurface.

Development geologists and engineers employ models to assist in the characterization of their reservoirs. These models take many forms, including the use of modern analogues of the reservoir's interpreted depositional system, outcrop analogues of a unit assumed to have formed under similar conditions, physical scale models of the depositional system, and numerical simulations of the reservoir built using mathematical short-cuts to simulate the physics of reservoir construction. Many published studies attest to the usefulness of such models, at least as providing first approximations of reservoir character, although it is almost always the case that discrepancies develop between the predicted character of the reservoir and the actual performance of the reservoir, as development proceeds (the issue of "history matching"). Several general studies of the modeling process have appeared in recent years, that have provided excellent introductions to the strengths and limitations of the various approaches (e.g., Alexander 1993; Bryant and Flint 1993; Geehan 1993; North 1996).

In a lengthy and thorough review of the area of modeling and prediction of subsurface fluvial reservoirs, North (1996) emphasized the complexity and variability of fluvial successions and the difficulties in predicting fluvial architecture in the subsurface. He discussed the various conceptual approaches that have been used to systematize our understanding of fluvial systems, including vertical-profile-based facies modeling, architectural-element analysis and sequence stratigraphy. He noted the problems caused by the simultaneous actions of the various autogenic and allogenic sedimentary controls. He demonstrated that limits of vertical seismic resolution and the limits imposed by a borehole network, even within a mature basin, may limit the ability of the geologist to accurately define and predict fluvial architecture with the quantitative rigour required by development engineers. Ethridge (2011) likewise, in an appraisal of the methods of sedimentological interpretation of ancient fluvial systems, reviewed the many attempts to classify fluvial channels and channel systems, pointing out the inconsistencies in terminology and the fact that such classifications have not, in fact, assisted greatly with the interpretation of the ancient record.

North (1996, p. 451) suggested that the computer models of flow in channels (as recently summarized by Bridge, 2003), which provide predictions of vertical profile and paleocurrent variations, are valuable, as providing the basis for more reliable reconstructions of channel form and style than earlier, descriptive models, but acknowledged that sufficient data would rarely be available from the subsurface to make this a practical tool. These numerical models are based on geomorphic data bases of channel dimensions, from which sets of equations have been derived that express the relationships between such parameters as channel width, depth, meander wavelength, discharge, etc. (e.g., Ethridge and Schumm 1978; Bridge and Mackey 1993b). North (1996, p. 452) noted the inadequacy of the data base on which paleohydraulic reconstructions have been based, the large errors inherent in the standard equations, and the procedural errors involved in using the output from one equation as the input for another. Many studies, including that of Bridge and Mackey (1993b), have addressed the issue of the paucity of data, but the conceptual question discussed by Alexander (1993) and Geehan (1993) remains: how do we know we are using the right analogue?

Weissmann et al. (2011) offered an even more fundamental criticism of the data base of fluvial studies on which modern fluvial sedimentology rests: they argued that most of the modern river systems, the descriptive features of which have been used to construct modern facies models, are located in degradational settings. They asserted that these studies are of limited relevance in the interpretation of ancient successions which, by their very existence, indicate the long-term persistence of aggradational environments. They stated (p. 330):

> We believe that these studies of fluvial systems in degradational settings have validity in terms of channel processes and products at the scale of bar forms, macroforms, and channel belts. However, they do not inform us about the way the macroform-scale deposits stack into overall 3D basin-fill architecture.

I address this argument in Sect. 7.3.2.

A theme throughout the discussions by North (1996) and the concluding remarks in the book of which that paper is a part (Carling and Dawson 1996) is the lack of information about modern rivers, a refrain expressed many times by J. S. Bridge, as well. For example, Mackey and Bridge (1995, p. 28) concluded that "There is a critical need for more comprehensive architectural data from modern fluvial systems, especially data related to processes controlling floodplain geometry and channel pattern over periods of thousands of years." They called for more comprehensive physical models of flow, sediment transport, channel geometry and the effects of tectonism and base level change. However, the usefulness of such models would still be questionable, for the reasons discussed below. North (1996, p. 399) noted that:

> The geological emphasis needed is often determined by the economic and engineering parameters of the project. So in a hydrocarbon reservoir analysis, for example, while the geologist may be fretting over the sinuosity of the ancient river, the engineer may be much more concerned by the impact on channel-sand permeability and porosity of the variations in diagenesis.

Tye (2004) argued that the documentation of surface form, without the need for subsurface analysis, could provide an invaluable input into reservoir studies by providing constraints on the scale, orientation and interrelationships between reservoir components, such as channels and bars, so long as the appropriate modern analogue had been selected from which modeling input data was derived. He illustrated his argument with examples of the use of measurements on selected modern rivers and deltas as input into an object-based three-dimensional reservoir model. He acknowledged, however, that his "geomorphology" approach could not take account of the erosional relationships between successive channel-belt units. This is where knowledge of the subsurface architecture must be added in.

The problem of documenting fluvial architectures from modern river systems has largely been solved by the development of ground-penetrating radar (GPR). This geophysical technique is superbly adapted to documenting the shallow subsurface, providing high-resolution architectural data that can be related precisely to the surface channel and bar morphology (e.g., excellent case studies were provided by Best et al. (2003), Lunt and Bridge (2004)). Both the value and the limitations of modern architectural studies using GPR are well illustrated by the detailed study of the Sagavanirktok gravelly braided river in Alaska by Lunt and Bridge (2004) and Lunt et al. (2004). These papers contain detailed documentation of the channel and bar architecture, documented with numerous GPR profiles. From the GPR data the authors extracted a set of "vertical logs of typical sequences through different parts of compound bar deposits and channel fills" (Lunt et al. 2004, Fig. 24d). They also developed a table relating "stratal thicknesses measured in boreholes" to the "widths of different scales of stratasets" (Lunt et al. 2004, p. 410 and Table 2.3). They stated that this "quantitative three-dimensional depositional model … will allow prediction of the dimensions and spatial distributions of different scales of stratification …" However, they then go on to say that "reconstructing the origin and evolution of compound bar deposits from only recent aerial photographs or cores is impossible. It is also impossible to determine from core whether a compound bar was a point bar or a braid bar." They also assembled some modern data relating to the width-depth relationships for the channel belt deposits of recent braided and meandering rivers and concluded that this ratio is widely variable and that there may be very little difference between the two river styles in terms of the channel-belt deposits currently accumulating.

Here, then, is the first of the two major problems with modern analogues for interpreting the ancient record: snapshots of a modern river (surface maps, aerial photographs) do not necessarily reveal the internal structure of the bars and channel deposits beneath the surface. For example, an apparently simple point bar in a braided system may, upon dissection or GPR surveying, reveal an internal structure partly composed of the remnants of a different type of bar, or of an earlier point bar with a different orientation, upon which the modern bar form has been superimposed by the latest configuration of the adjacent active meander bend. Best et al. (2003) documented the evolution of a single large braid bar in the Jamuna (Brahmaputra) River in Bangladesh. This bar, 1.5 km long in a downstream direction, migrated downstream a distance equal to its own length in a little over a year, and temporarily doubled in downstream length. How relevant to the study of the ancient record is the detailed documentation of such an ephemeral feature, other than to illustrate short-term bar-forming processes? How much of this bar is likely to make it into the preserved record?

In its simplest condition, the evolution of a braided channel can be considered as the development of opposite-facing low-sinuosity meanders migrating away from a central (mid-channel) bar (Bridge 1993). The work of Ashworth et al. (2000) explicitly ruled out this mode of evolution in the case of the bar they studied, although they made a comparison with the small bar in the Calamus River, Nebraska, analyzed by Bridge et al. (1998), which the latter demonstrated to have grown by a comparable pattern of lateral and downstream accretion from an upstream nucleus. Where bar migration is symmetrical, as proposed by Bridge (1993), channel scour would be expected to sweep out an erosional channel form approximating the width of two channels plus the intervening bar. Assuming two channels of second-order Brahmaputra scale (in the terminology of Bristow (1987)), each 2 km wide, and a mid-channel bar also 2 km wide, if both channels were filled prior to abandonment this could theoretically generate a second-order sand body bounded by a fifth-order surface (the numbering refers to the channel-scale bounding surface classification of Miall (1988, 1996, 2010a)) in the order of 6 km wide. With an average depth of 12 m such a sand body would have a W/D ratio of 500. However, this scenario is quite speculative. Several groups of researchers have demonstrated patterns of active anabranch migration and bar growth and erosion in the Brahmaputra/Jamuna River (Thorne et al. 1993; Ashworth et al. 2000), which indicate that sand bodies of the full theoretical width estimated here may never develop. Sand bodies bounded by surfaces of fifth-order rank are likely to be substantially less than 6 km wide. The final preserved architecture of sand bodies of the type described by Ashworth et al. (2000) would depend on the balance between (1) lateral growth of the bar under conditions of anabranch migration, and (2A) erosional incision brought about by events of avulsive anabranch switching, or (2B) migration and lateral erosion of an anabranch from another location within the channel belt. Final preserved sand body widths are presumably somewhere between the hypothetical maximum of 6 km and the width of individual bars—a minimum of 1 km. How useful are estimates with such wide error margins? I return to this question in Sect. 7.4, where the Brahmaputra/Jamuna River is discussed as a possible analog for the interpretation of the Hawkesbury Sandstone, Australia.

The second of the major problems is that well data (including core logs) relating to the internal architecture may be as poor a guide as surface form as a diagnostic tool for reservoir body evaluation. Lunt et al. (2004) reconfirmed the point argued many years ago (e.g., see Miall 1980; Collinson 1986) that vertical profiles are not reliably diagnostic of fluvial style, let alone of bar character within a river of known style. Even with a detailed core record it may be difficult to impossible to determine whether a particular vertical profile relates to a single channel-fill record or to superimposed fragments of several or many channel and bar deposits, such as the one documented by Best et al. (2003). Interpretations derived from core should therefore include the development of several alternative scenarios for further testing.

The demonstration of statistical relationships between channel thickness and width may be useful for characterizing individual rivers, but such relationships should be used with great caution in examining the ancient record. The problem is that even detailed GPR documentation of a modern river system relates only to the present-day snapshot of the deposits. On the short term (decades to hundreds of years) the architecture relates to the preservation of fragments of bars and channels formed, modified and eroded under the existing channel pattern. But none of this present-day deposit has yet made it into the geological record (this is, in part, what Weissmann et al. objected to, as noted above). On the longer term (from thousands of years up to geological time scales) the pattern of preservation is influenced by subsidence rates and climate change. In addition to the fragmenting of channels and bars within the short-term time frame of channel migration and avulsion there may be erosional incision caused by channel systems at much later time periods, which may partially or completely remove the earlier deposits and which may demonstrate different styles because of changes in long-term allogenic controls. Given slow subsidence rates it is quite conceivable that a given stratigraphic unit could contain the amalgamated, mutually incised fragmentary deposits of different river styles that were active tens to hundreds of thousands of years apart and which could have generated channel and bar deposits with significantly different internal character and thickness-depth relationships (e.g., see Blum and Törnqvist 2000; Ethridge and Schumm 2007; Sheets et al. 2008). In Chap. 6 we address the issue of the relationship between alluvial architecture and accommodation generation.

Shanley (2004, pp. 171–172) argued that although much geomorphic information is available from studies of modern rivers, "the interplay of subsidence, base level, and magnitude of sediment supply exerts a far greater control on the degree to which fluvial [channel] deposits are amalgamated or isolated than the many short-term processes commonly viewed in the study of modern analogs." Gibling (2006) has documented with a thoroughness not previously attempted the enormous range in the dimensions of channel bodies in the modern and ancient record, the variability in sedimentary controls, and the difficulties inherent in interpreting and modeling fluvial systems from limited data. As Ethridge and Schumm (2007) noted: "Because several controls can produce the same effect (convergence) and one control may produce different effects (divergence), unambiguous interpretations [of the ancient record] are not possible."

Given the normal variability of geological processes, the assumption of architectural complexity and variability should be the null hypothesis for the purpose of exploration and development. For these reasons, it is suggested that the statistical relationships developed for reservoir body dimensions and the numerical models that are based on them (e.g., Bridge and Mackey 1993a, b; Mackey and Bridge 1995) are most appropriately used only in a very preliminary fashion as guides to the development of several alternative scenarios for reservoir interpretation and development. Shanley (2004) demonstrated this approach with the use of an array of different equations for the estimation of sand-body widths from log- and core-derived thickness data.

Modern sedimentological interpretations began in the 1950 s with the recognition of the value of the "vertical profile" as a diagnostic feature of depositional environment, a development largely attributable to the work of Esso and Shell Development geologists, who recognized the repeated nature of certain profiles on wireline logs (Nanz 1954) and compared these to profiles obtained from the study of selected modern environments, including fluvial point bars (Bernard et al. 1962). At about the same time, Allen (1963, 1964, 1965a, b) working largely in the Devonian Old Red Sandstone of the Anglo-Welsh borderlands area, began to establish the link between meander migration, point bar formation, and the relationship between width, depth and other channel attributes as preserved in the rock record. Leeder (1973) noted a useful relationship between the geometry of point bar deposits and the dimensions of meandering channels. Geomorphologists, such as S. A. Schumm, provided much food for thought from their study of modern river systems (e.g., Schumm, 1977) and the several generations of numerical models that have been developed, most recently (Bridge and Mackey 1993a, b; Mackey and Bridge 1995) have built on all this earlier work to simulate alluvial architectures based on selected input data and sets of geomorphic equations based largely on observations of the fluvial styles of modern rivers. This history (up to the mid 1990s) is recounted in some detail in the history chapter of "*The geology of fluvial deposits*" (Miall 1996, Chap. 2).

Amongst the foundational work necessary for modeling have been attempts to document and categorize fluvial deposits based on their interpreted fluvial style, major milestones in this progress being the papers of Fielding and Crane (1987) and Robinson and McCabe (1997) (e.g., see Fig. 2.8), and culminating in the authoritative compilation by Gibling (2006), the last word in empirical data collection on the size and shape of all types of preserved sandstone and conglomerate body in the ancient record. The hope has been that from all this generalization would emerge patterns that would enable reservoir geologists to take the very few bits of information that are normally available from subsurface exploration, such

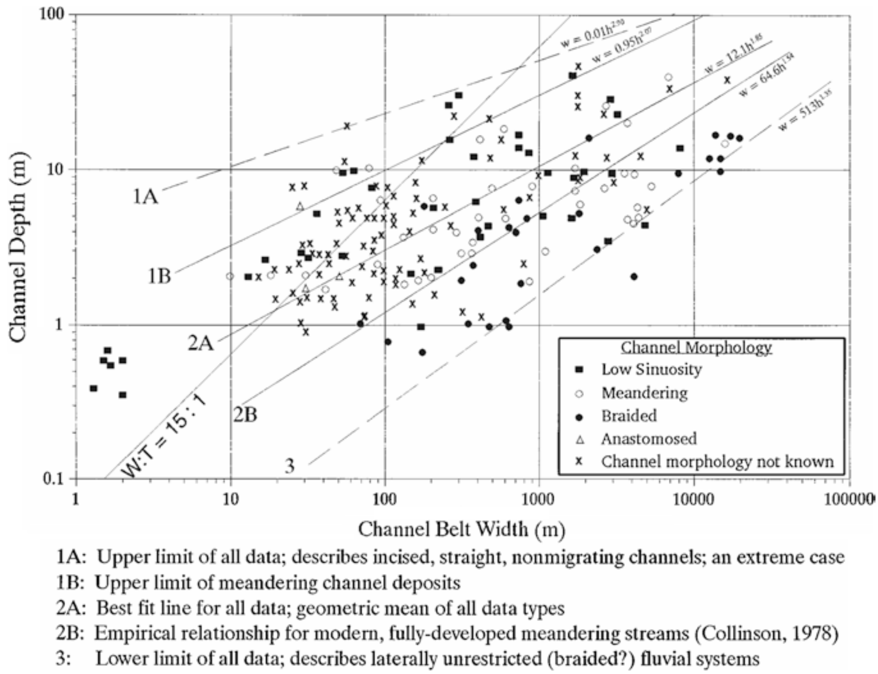


Fig. 2.8 The relationship between channel width and depth for various fluvial styles. Adapted by Robinson and McCabe (1997) from an earlier synthesis by Fielding and Crane (1987). AAPG © 1997. Reprinted by permission of the AAPG whose permission is required for further use

as sand-body thickness and lateral extent (based on sometimes questionable stratigraphic correlation exercises), and from these develop reservoir models that could be simply handed over as end products to the production engineer.

The effort, which has now been underway for more than half a century, to document and categorize fluvial facies models is still not complete. Miall (1985) summarized architectural work that had led to the recognition of 12 distinctive styles, and this was later (Miall 1996, Chap. 8) expanded to 16. Long (2011) succeeded in identifying examples of most of these in the Precambrian and Early Paleozoic rock record. Nonetheless, some workers have argued that additional models are still necessary. Fielding et al. (2009, 2011) defined a new model for tropical rivers characterized by seasonal, semiarid to subhumid conditions, and applied this model to an interpretation of the Upper Paleozoic record of Atlantic Canada (Allen et al. 2011). Such criteria as "Sandstone bodies with complex and abrupt lateral variations of sandstone and pedogenically modified mudstone" or "Paucity of lateral or downstream accreting macroforms" are cited as features that characterize high-discharge rivers in such settings. However, as noted in Sect. 5.2.2, the climatic interpretation in this study depended largely on paleobotanical and paleosoil evidence. Most of the facies and architectural features that are asserted to be characteristic of the sandstones formed in this climatic setting are common in sandstones deposited in many depositional settings, and it seems very unlikely that the climatic setting of a fluvial deposits could be unambiguously interpreted based solely on the lithofacies assemblage or architecture of the clastic components of the succession.

On the other hand, North and Davidson (2012) pointed out a number of misconceptions in the use of terms relating to unconfined flow and the resulting deposits. They demonstrated that such terms as "sheetflood" and "sandflat" are poorly defined and have been used in incorrect ways through much of the sedimentological literature. This has important implications for interpretations of the subsurface. For example, fluvial deposits characterized by a predominance of plane bedding (lithofacies Sh, architectural element LS of Miall (1985)) have in many cases been described as the product of sheetfloods. The implication is that plane beds develop beneath bodies of water than may be described as sheet-like in geometry—lacking bedforms—but the interpretation commonly includes the implication that such flow conditions are most characteristic of high-discharge events that overtop river banks and spread out onto the floodplain as a fluvial "sheet". The condition of high-discharge sedimentation across the floodplain is well described by the term "unconfined flow", but this does not necessarily imply the plane-bed flow-regime condition. Indeed, unconfined flow may include a wide range of deposits and, conversely, the plane-bed condition may be developed in channelized flow and has recently been cited (Allen et al. 2011) as one of the characteristics of seasonal tropical rivers. Given that the term "sheetflood" carries definite implications as to geometry—a criterion of key importance to the reservoir geologists, such distinctions in terminology are of more than academic importance. As North and Davidson (2012) note, the term "sandflat" is even more poorly defined.

Increasing knowledge of the variety of fluvial depositional environments is leading to a re-evaluation of some earlier facies interpretations. Even the famous "fining-upward cycles" of the Old Red Sandstone of Britain are falling victim to this phenomenon. These cycles, as exposed along coastal cliffs in South Wales, were amongst the first to be interpreted as the product of point-bar sedimentation (Allen 1963b). An increase in our knowledge of dryland environments, particularly the Eyre Basin of interior Australia, has led to a reinterpretation of exposures of these rocks in Pembrokeshire, in South Wales, as the deposits of ephemeral systems, in which lateral point-bar migration comprised a very minor component (Marriott et al. 2005).

2.3.2 Facies Models and the Subsurface

Has the work of facies analysis done what it set out to do—assist the subsurface geologist to map and assess the reservoir potential of fluvial sandstone and conglomerate bodies? After a half century of research the answer has to be, not really.

Consider Figs. 2.9 and 2.10. Figure 2.9 established three basic models for developing reservoir simulations. The layer-cake model is one that might be

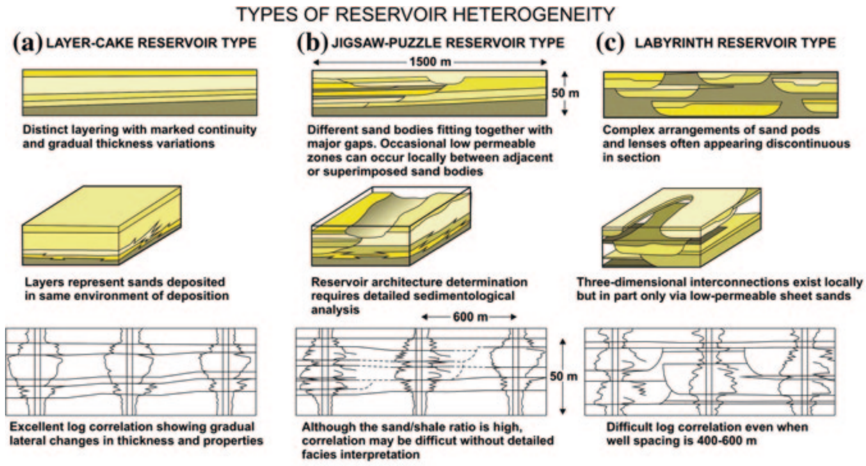


Fig. 2.9 Three reservoir geometry models (after Weber and van Geuns 1990)

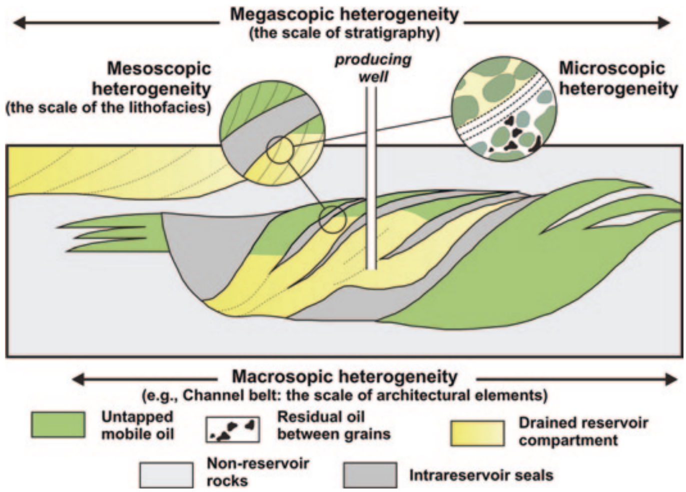


Fig. 2.10 Four scales of reservoir heterogeneity (after Tyler and Finley 1991)

expected to be demonstrated by such depositional systems as sheet-turbidites within submarine down-fan settings. The jigsaw-puzzle and labyrinthine models are characteristic of many depositional systems and are difficult to map and from which to develop useful predictions. With the possible exception of deposits composed entirely of laminated sandstone sheets (architectural element LS; the "flashy, ephemeral, sheetflood, sand-bed river" of Miall (1996), Sect. 8.2.17), most fluvial systems may be characterized by one or other of the jigsaw-puzzle or labyrinthine models. Much of the sedimentologic research into fluvial systems during the last five decades has been devoted to attempts to provide tools for the estimation of the sizes, shapes, interconnectedness and orientation characteristics of the complex types of reservoir body to be expected in reservoirs that may be described by the jigsaw and labyrinthine models.

Figure 2.10 illustrates how reservoir complexity may be considered on at least four different scales. Techniques for mapping and prediction vary between these scales. The simplest to document are the largest and the smallest, the largest because the scale of megascopic heterogeneity may be expected to exceed that of well spacing at field development stage, and the smallest because this is the scale that may be reliably documented from well cuttings and the thin-sections made from them. Mesoscopic and macroscopic heterogeneity are sedimentologic in nature and, in the case of fluvial systems, reflect the size and architecture of channel systems and their constituent architectural elements. Tyler and Finley (1991) suggested that a knowledge of the heterogeneities at these intermediate scales could increase production efficiencies dramatically, by providing guidance for careful placement of infill drilling or horizontal production wells during enhanced recovery programs. However, they noted that "mobile-oil recovery is inefficient in highly channelized reservoirs." Much of the intermediate heterogeneity may border on the unresolvable. This is why considerable ingenuity has been devoted to the development of numerical models for fluvial architecture, based on statistical probabilities, as summarized above. Advanced sedimentologic research is now, in practice, aimed more at refining the data base for statistical modeling than for documenting the actual specifics of individual reservoirs.

A large part of the problem is the naturally occurring inconsistency of fluvial systems. Channel morphology changes downstream in response to changes in valley slope, sediment load, bank materials, climate, or tectonic regime (e.g., see Schumm 1977), and the same controls may cause changes through time in the morphology of a particular river reach. It is therefore unwise to assume that fluvial style will remain constant throughout a given stratigraphic unit. This point is particularly relevant to the case of the largest river systems, and is examined further in Chap. 7.

Figure 2.11 illustrates a typical example of a large modern system, part of the Congo River and some of its tributaries. The four major rivers visible in this image display at least three distinct styles, each reflecting the nature of upstream and local controls on fluvial magnitude and discharge variability, sediment load, bank composition and vegetation cover. Each of the rivers exhibits a moderate degree of variability along its length. The natural world is full of examples of this type, where the basin centre and the various watersheds bordering it are characterized by different source-area geologies and microclimates, leading to great within-basin variability in fluvial style. Now imagine an ancient fluvial deposit in a major basin developed by such a complex of rivers. Attempts to develop a geostatistical description of each river might have some success if it was known in advance where each river was located, but this, of course, begs the question. Generalizations for the whole basin from the data available from a few dozen exploration wells—the most likely available at an initial discovery phase, would be hopelessly inaccurate. Diagrams provided by Martin (1993) illustrate the problem (Fig. 2.12). The value of these diagrams is that they place a basic geological problem in a

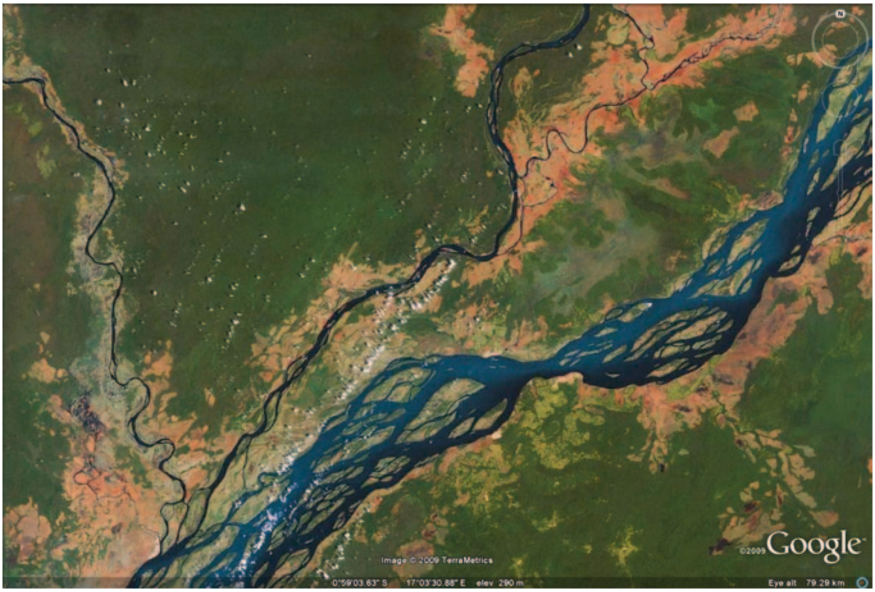


Fig. 2.11 Part of the Congo River system. The image is about 40 km from east to west (Image reproduced from Google Earth. Terra Metrics © 2009)

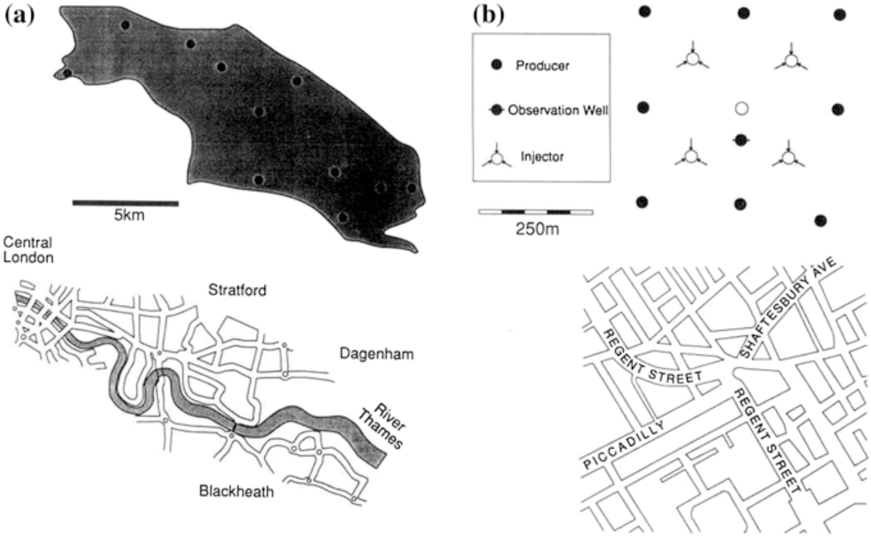


Fig. 2.12 Comparison of common development problems with recognizable scales of familiar human structures—a necessary step required to enable appropriate judgments about well spacings and the real scales of depositional systems. **a**. Location and spacing of appraisal wells of the Snorre field, offshore Norway compared with the major roads and Thames River of east London (Martin 1993, p. 340, Fig. 3); **b**. Well locations of an enhanced recovery pilot project superimposed on the detailed street plan of the Piccdilly Circus area of London (Martin 1993, p. 341, Fig. 4)

recognizable human-scale context. It is commonly far too difficult to make the necessary scale comparison between a poorly known basin and actual depositional systems. Diagrams like this are a great help. More on this topic in Sect. 7.3.2.

Some river systems, and their deposits, are, hopefully, described as “sheet-like” in character. This was a term proposed by Miall (1996, p. 484) for a category of reservoir units deposited by “steep-gradient, bed-load systems, such as braided rivers, where channels comb across broad areas of the valley floor.” Much of the main reservoir unit in the Prudhoe Bay field (Sadlerochit Formation) has been described using this term. For example, Martin (1993, p. 335) cited the Prudhoe Bay Field as an example of a sheet-like reservoir with high net/gross ratios, porosity and permeability and with oil recovery factors commonly up to more than 50 %. Reservoirs are said to be in internal pressure communication with common field-wide contacts. The gravel-braided rivers of the modern Canterbury Plain, South Island, New Zealand (Fig. 2.13, may be considered a modern analogue. However, already by 1989 the Prudhoe Bay field was showing signs of troubling internal inconsistency, as evidenced by the fact that the production team in Anchorage was interested in the fluvial architectural work that was being developed at the time by myself and others (e.g., Miall 1988). Once depletion sets in and reservoir pressures drop, minor internal barriers and baffles to flow become more significant, and production characteristics become more unpredictable (Fig. 2.14).

Fig. 2.13 Gravel-bed braided rivers of the Canterbury Plains, South Island, New Zealand

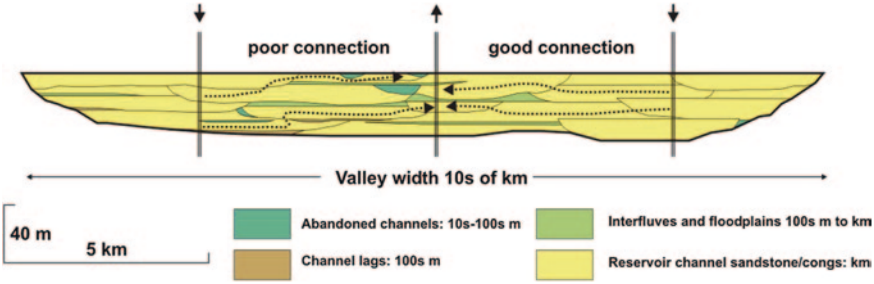


Fig. 2.14 How small variations in the composition and architecture of a channel belt can affect reservoir performance. To the *right*, the succession compares to the "jigsaw-puzzle" model of Weber and van Geuns (1990); to the *left*, a comparison may be made to the more complex "labyrinthine" model (see Fig. 2.1)

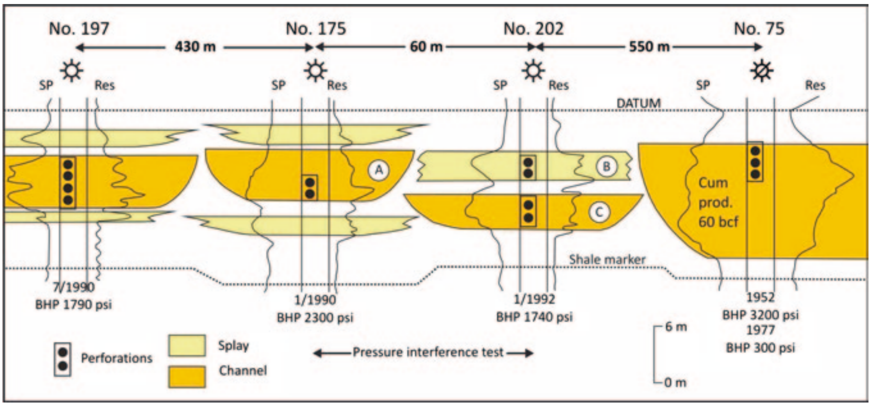


Fig. 2.15 An extreme example of reservoir compartmentalization. See text for explanation (Hardage 2010)

An extreme case of reservoir heterogeneity was described by Hardage (2010). The problem is illustrated in Fig. 2.15. It had been assumed that the reservoir sand "A" in hole 175 was in communication with sand bodies in the adjacent holes. However, repeated pressure tests demonstrated that this was not the case. After packing hole 202 to isolate sand "B", a pressure pulse was run in hole 175, but was not "felt" at all by pressure changes in hole 202. The same result was obtained when packing was used to isolate sand "C." The conclusion, that fluid communication could not be assumed even over a distance of 60 m, was troubling in the particular case, and provides a warning against relying on simplistic stratigraphic and architectural reconstructions.

Figure 2.16 illustrates the general problem, one discussed briefly in an earlier paper (Miall 2006a). It is a common complaint (e.g., Bridge and Tye 2000, p. 2006) that facies models for fluvial deposits are of limited use in interpreting the depositional setting of

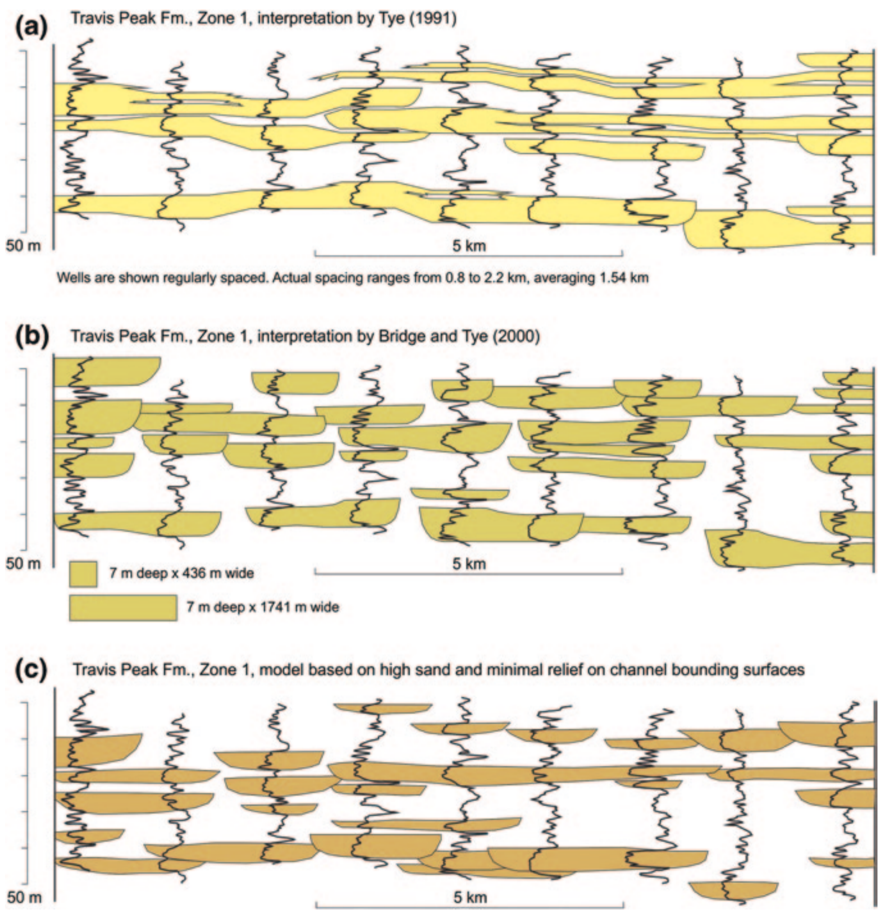


Fig. 2.16 Three interpretations of the braided-fluvial deposits of the Travis Peak Formation, Zone 1 (Early Cretaceous, East Texas). **a** Initial interpretation, by Tye (1991), based on detailed core and isopach mapping study. Arbitrary equal well-spacing is used in this and the subsequent diagrams. **b** A reinterpretation by Bridge and Tye (2000), based on assumptions of narrower channel belts. Rectangular boxes at the base of this panel indicate range of channel-belt sizes predicted from estimated bankfull depth, using the equations of Bridge and Mackey (1993b). Their own model, shown here, does not make use of this range of values;.**c**. An alternative model developed by the present writer (Miall 2006a), based on two basic guidelines for interpreting petrophysical logs: (**a**) channels normally have flat bases, and (**b**) the main sand bodies are indicated only by blocky-shaped, low-value gamma ray signatures

reservoir bodies because they are incomplete or misleading. In fact, all that such models were ever intended for was general guidance, to serve as "norms" and "predictors," to use Walker's (1976) terms. Bridge (2003, p. 222) correctly stated that channel patterns cannot be deduced from vertical profile data alone, thereby confirming earlier observations by others who also dealt with the ambiguities of vertical profile data (e.g., Miall 1980, 1985, p. 263; 1996, p. 38–42; Collinson 1986, p. 59–60; Shanley 2004).

While noting the inadequacy of vertical profile data, the suggested solutions are in fact variations on this same approach. Thus, Leclair and Bridge (2001) explored the relationship between crossbed thickness and bedform height so that the known dependence of bedform height on flow depth may be used to estimate channel depth. Use of this relationship for subsurface analysis depends on being able to obtain useful information about crossbed thickness from vertical profile data. Another example of the dependence on vertical profile data is the subsurface methodology proposed by Bridge and Tye (2000) in which diagrams that they explicitly label as "idealized vertical sequences of lithofacies and wireline-log response" are offered as improved tools for interpreting channel geometry and width.

Bridge and Tye (2000, p. 1223) claimed to have offered a "fresh approach" to the quantitative evaluation of subsurface fluvial architecture (Fig. 2.16). This approach has four components, about which the following may be said:

(1) Regarding their "new models for the three-dimensional variation of lithofacies and petrophysical log response of river-channel deposits." These models are descriptions of bar (macroform) growth and migration. They are based on the long-standing idea that they are independent of channel planform (e.g., Allen 1983; Miall 1985) and a growing conviction that such processes are scale-independent (Sambrook Smith et al. 2005). In that sense the models are not new. They incorporate much new data derived from studies of modern rivers and ancient analogues, but they do not deal with such well-known fluvial processes as crevassing, and the scouring that takes place at channel confluences, both of which can generate distinctive facies architectures. Nor do these models take into account the issues of preservability of channels and their individual elements.

    Elsewhere Bridge (2003, p. 223) stated "When attempting to reconstruct paleochannel patterns from ancient deposits it should be realized that channel patterns in a particular reach of a channel belt can vary markedly in space and time. This may be due, for example, to local variations in bank materials, localized tectonism, the effects of particularly severe floods, or bed cut-offs." The value of theoretical models is therefore moot.

(2) "Distinction between single and superimposed channel bars, channels and channel belts." It is asserted, but not demonstrated, that such distinction may be made. In fact, as has long been known, distinction between these three scales is difficult to impossible in core because of the lack of uniqueness of any of the defining characteristics of the deposits (e.g., Miall 1980, 1988). Figure 2.17 illustrates the four major scales of "fining-upward cycle" that may be observed within fluvial deposits. Distinction between them based on limited drill-hole data is likely to be quite difficult.

    With the possible exception of the scale (thickness) of individual crossbed sets, none of the features of vertical profiles that are observable in core, including vertical succession and the nature of bounding surfaces, is amenable to unique interpretations. Application of the "new models" offered in this paper suffers from the problems of fragmentary preservation, which is all too common in

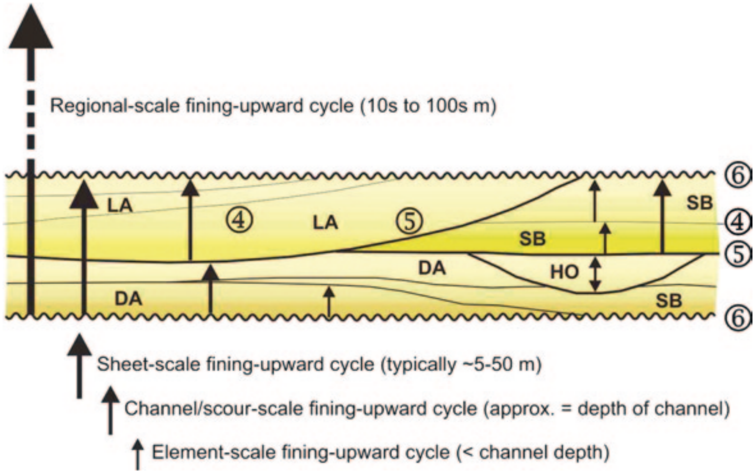


Fig. 2.17 The four scales of fining-upward cycle commonly observed in ancient fluvial deposits. Adapted from Godin (1991)

fluvial systems. Lunt et al. (2004) specifically acknowledged this problem in a related paper in which they develop a gravelly-braided fluvial model.

(3) "Interpretation of maximum paleochannel depth from the thickness of channel bars and from the thickness of sets of cross-strata formed by dunes". The interpretation of channel bar deposits is affected by the considerations noted in the previous paragraphs. Estimates made from crossbed thickness may be more reliable, but leave open the question of their representativeness. For example, deposits formed following deep scour may be more preservable than those that form during "normal" conditions, and are likely to be larger and thicker than average, but how representative are they?

(4) "Evaluation of methods for estimation of widths of sandstone-conglomerate bodies that represent either single or connected channel belts." No new ideas are in fact offered here. The reader is referred to new equations relating thickness to width that are claimed to be "more generally valid because they are based on broader data sets than previous equations or on theoretical principles." But no single data set can account for the simultaneous variations in subsidence rate, sediment supply, and discharge that characterize natural fluvial systems. The Bridge and Tye (2000) study made use of the empirical equations of Bridge and Mackey (1993b) to estimate channel-belt width. These equations are based on unspecified data bases that presumably incorporate many different types of river, but given the variability in fluvial form and the geological variability in the processes that govern fluvial style, no objective reason can be provided for preferring one equation over another. The immense natural variability in form and scale is well documented by Gibling (2006).

Bridge and Tye (2000) developed a new interpretation of a sandstone unit that had earlier been described by Tye (1991). The original 1991 interpretation is shown in Fig. 2.16a.

In fact, the reinterpreted channel belt model shown in Fig. 2.16b (redrawn from Bridge and Tye (2000), Fig. 2.9c) does not correspond to the dimensions calculated from their new equations. Bridge and Tye (2000, p. 1220) stated "If maximum bankfull flow depth in the Travis Peak Formation [estimated from core] ranges from 6 to 10 m, mean bankfull flow depth is 3–5 m, and the range of channel-belt width is predicted to be 436–1741 m using the empirical equations from Bridge and Mackey (1993b)." Two scaled rectangles, with these dimensions, are shown in Fig. 2.16b. According to these estimates, but not as shown in their diagram, most of the sand bodies would not be intersected by more than a single well, and sand body interconnectedness would be very low, unless there are many more similarly narrow sand bodies between and not intersected by any of the wells. No particular value is attached to the third model shown here (Fig. 2.16c). It was drawn by this writer to be as faithful to the logs as possible, and indicates a possible zone of well-connected sand bodies near the centre of the section, and a low-sand interval at the top. As discussed throughout this book, real fluvial systems, as opposed to numerically simulated models, may be highly variable. Only surveillance methods (seismic time-slices, 4-D seismic, pressure tests) could determine the relative "truthfulness" of these models.

Most attempts to describe and predict fluvial reservoirs based on geological data have made use of outcrop analogue data. Bridge and Tye (2000, p. 1217) argued that outcrop ancient-record analogues for subsurface comparisons are rarely adequate because of a lack of fully three-dimensional data and uncertainties about the appropriateness of the analogue being used for each specific case. In some projects, one or more specific outcrop case studies are referred to; in other cases use is made of existing statistical relationships for relating to each other the various scale parameters in fluvial systems. Various statistical techniques may be referred to, or numerical modeling of the system attempted. But however sophisticated the statistics and the numerical model, ultimately these projects must resort to some means of determining appropriate input data from the real world of actual fluvial systems.

One of the most detailed studies of this type was the thesis work by Martinius (1996; see also Martinius (2000)) who derived quantitative sand body, petrological and petrophysical data from two outcrop studies of Tertiary units in Spain. The use of detailed sedimentological studies in a mature field was described by Tye et al. (1999). Their work on the Ivishak Formation in the Prudhoe Bay field showed that production surveillance data could be used to refine the prevailing sedimentological model and the enhanced recovery design, with subsequent improvements in history matching. Willis and White (2000) provided a very detailed outcrop study of a tidally-influenced delta deposit in Wyoming from which they developed probability scale distributions for five distinct facies types, and then carried out flow simulations. Karssenberg et al. (2001) attempted to demonstrate the utility of the three-dimensional numerical model of Mackey and Bridge (1995) by "conditioning" the model with data from five synthetic wells to generate a realistic simulation. Yu et al. (2002) studied a large outcrop of a Jurassic fluvial system in China and developed from this some generalizations about fluvial architecture and petrophysics that they offered as an analogue for interpreting producing reservoirs in east China. Svanes et al. (2004) defined “genetic types” of sedimentological objects in vertical profile, and used these in conjunction with 3-D seismic data to develop a fluid drainage model in a producing field. They pointed out the difficulties in making adjustments to a stochastic reservoir model to accommodate new input from well data or surveillance data (the “conditioning problem”).

Sand body architecture is a product of fluvial style and accumulation processes. In other words, it’s all about the nature of the surface fluvial system—the size, shape and orientation of channels and their component architectural elements (including bars and crevasse splays)—and about how these fluvial systems behave over time—the nature of lateral channel movement and rates of subsidence. In the next section we examine the issue of fluvial style. In Chap. 3 we discuss autogenic controls on the accumulation of fluvial systems over time, with a particular focus on the process of avulsion. With this information in mind, we can then focus on the issue of sand body architecture. Section 3.7 reviews modeling work in this area that has been carried out with a view to understanding sand body connectedness—the key issue in maximizing the productivity of fluvial reservoirs.

2.3.3 Controls of Fluvial Style

One of the major problems with the study of fluvial style is that the terms used to describe style are not mutually exclusive; indeed they describe different conditions and different process entirely. The term *meandering*, originally derived from the name of the Buruk Menderes River in Turkey, refers to the pattern of sinuous river bends that characterizes many rivers, especially (but not exclusively) those carrying a relatively fine-grained bedload or suspended sediment load. (The Buruk Menderes River is now in effect an underfit stream because of large-scale water diversion for irrigation. Classic meandering channel and point-bar surface form representing pre-modern fluvial architectures is visible as textural patterns in aerial imagery of the floodplain, [e.g., on Google Earth] which has now been entirely developed for mixed-crop farming). The term *braiding* refers to a pattern of multiple channels separated by bars and temporary islands. For many years the term *anastomosing* was considered to be synonymous with the term braiding, but it is now recommended that the term be restricted to rivers characterized by a network of stable channels of low- to high-sinuosity (Miall 1996, p. 15). Nanson and Knighton (1996) and Knighton (1998) use the term *anabranching* as a catch-all term for a range of similar channel styles, of which the anastomosing style of Smith and Smith (1980) is the most well known. In contrast to braided rivers, anastomosed rivers are typically characterized by stable, vegetated floodplains. Braiding and meandering describe processes that can occur at the same time in the same river, with the result that some rivers can be described using both terms, and this is one of the reasons why attempts by geomorphologists and geologists to classify and explain fluvial styles are still in a state of flux.

How did these terms develop, what do they tell us and, following on from that, how does a determination of fluvial style assist with the problem of subsurface mapping?

Much of the early history of development of ideas about fluvial style, including the work of Davis, Chamberlin and others, has been summarized elsewhere (Miall 1996, Chap. 2). The various classification systems of Schumm and others are also discussed in that chapter. In the discussion that follows here an attempt is made to point out the confusions that are still present in much modern analyses, with the aim of arriving at some concepts and ideas that are useful to the geologist working with the ancient record.

Friedkin (1945) was probably the first worker to examine the issue of fluvial style systematically. He carried out a much-cited series of large-scale experiments to model meandering and braiding in a large flume. He attributed meandering to bank erosion, but did not explain why this occurs and why it is characterized by regularity. He agreed with the term "overloaded river" for braided rivers (Friedkin 1945, p. 16) and attributed the braided character to bank erosion and bedload deposition enhanced by easily eroded bank materials.

Leopold and Wolman (1957) demonstrated that at least nine variables interact to determine the nature of the resulting stream channel. They include discharge (amount and variability), sediment load (amount and grain size), width, depth, velocity, slope and bed roughness. Schumm (1968a) later showed that the amount and type of vegetation growth also will affect stream type and, therefore, climatic and geological factors must also be considered. It is still not possible to define the ranges of values that will invariably produce ariver of a given type although, as noted below, certain interrelationships between the variables are now well enough understood for some generalizations to be made. Leopold and Wolman (1957, pp. 72–73) stated:

> Channel patterns, braided, meandering, and straight, each occurs in nature throughout the whole range of possible discharges. Some of the largest rivers in the world are braided; for example, the lower Ganges and Amazon. More are meandering, of which the lower Mississippi is the best known example. Meanders are common in very small creeks and braids are common in many small ephemeral streams… [It has been observed that] a given channel can change in a short distance from a braid to a meander or vice versa, that the divided channels of a braid may meander, and that a meandering tributary may join a braided master stream. Such changes in a given channel or such different channels in juxtaposition can be attributed to variations in locally independent factors.

In what was for many years the standard textbook on fluvial geomorphology, Leopold et al. (1964, p. 281) made this general statement regarding river styles:

> River patterns represent an additional mechanism of channel adjustment which is tied to channel gradient and cross section. The pattern itself affects the resistance to flow, and the existence of one or another pattern is closely related to the amount and character of the available sediment and to the quantity and variability of the discharge. … separation of distinctive patterns is somewhat arbitrary.

Leopold and Wolman (1957) and Leopold et al. (1964, pp. 284–288) reported on a by-now famous experiment to simulate mid-channel braid-bar formation (a process summarized by Miall (1977, pp. 12–14)). They stated (Leopold and Wolman 1957, p. 50): “Braiding is developed by sorting as the stream leaves behind those sizes of the load which it is incompetent to handle… if the stream is competent to move all sizes comprising the load but is unable to move the total quantity provided to it, then aggradation may take place without braiding”.

Leopold et al. (1964, p. 282) discussed braiding in this way:

> Braided or anastomosing channels are often but not always associated with sandy or friable bank materials. Also, vegetation has similar effects; a change from non-braided to braided character is sometimes associated with a change from dense vegetation along the channel banks to sparse or no vegetation. Whether these coincident changes are causally related cannot usually be ascertained, although the coincidence is suggestive.

Note the by-now abandoned use of the term anastomosing in this quote.

Leopold et al. (1964, pp. 292–295) were very clear about the relationship between sediment load and fluvial style:

> Although the channels may meander at low stages, at overbank flow the braided river often moves nearly straight down its valley. … when two rivers of a given size of river (same discharge) are compared, braided channels occur on steeper slopes than meanders. Steeper slopes contribute to sediment transport and to bank erosion and are often associated with coarse heterogeneous materials. All these are conditions which contribute to braiding.
>
> Where coarse material is available, braiding may result from the selective deposition of the coarser material, causing formation of a central bar and thus diverting the flow and increasing erosional attack on the banks. This was observed in the flume, in the gravelly channels studied by us in Wyoming, and in braided proglacial rivers described by Fahnestock (1963) and others. Even in fine material, however, irregular deposition of bars and bank erosion may produce a braided pattern. The shifting channel may move gradually during low flows, but during floods major changes in the position of the thalweg can be produced. Because deposition is essential to formation of the characteristic braided pattern, it is clear that sediment transport is essential to braiding. It is also evident, however, that if the banks were unerodible and the channel width confined, the capacity of the reach for the transport of sediment would be increased, reducing the likelihood of deposition. In addition, any bars which formed would be removed as flow increased, since bank erosion could not take place. Thus, for the bars to become stable and divert the flow, the banks must be sufficiently erodible so that they rather than the incipient bar give way as the flow is diverted around the depositing bar. Sediment transport and a low threshold of bank erosion provide the essential conditions of braiding. Rapidly fluctuating changes in stage contribute to the instability of the transport regime and to erosion of the banks; hence they also provide a contributory but not essential element of the braiding environment. Heterogeneity of the bed material in the same way creates irregularities in the movement of sediment and thus also may contribute to braiding.

Schumm (1977, p. 106) similarly emphasized issues of sediment load and discharge variability in the development of the braided pattern:

> Although the records are short, they indicate that rivers with high ratios of peak to mean discharge are morphologically different from rivers with low ratios, In a general way this is substantiated … for two rivers in Jamaica. In this case the only factor that can explain the difference between the braided Yallahs River and the narrower, more sinuous Buff Bay River is the marked seasonality of precipitation in the Yallahs River drainage basin. Annual precipitation is similar in both drainage basins, but larger floods occur in the

> braided Yallahs channel. There have not been systematic studies of the influence of flood peaks or of the ratio of peak to mean discharge on channel morphology, but this is an area that warrants further attention.

And again, on p. 108:

> In addition to the size of the sediment transported, the relative amounts of bed load and suspended load also significantly influence themorphology of sand-bed streams. For example, along the Smoky Hill-Kansas River system in Kansas, discharge increases in a downstream direction, but channel width decreases from about 300 feet to less than 100 feet in central Kansas. Farther east there is a marked increase in channel width. These and other changes are attributed to changes in the type of sediment load introduced by major tributary streams (Schumm 1968b). Tributaries introduce large suspended-sediment loads where the width decreases, and large bed-loads or sand loads are added where width increases.

Elsewhere, Schumm (1977, p. 121) noted the importance of vegetation on the channel banks in stabilizing the channel. The action of depositing the coarser bed-load initiates mid-channel bar formation. In rivers of highly variable discharge competency will be similarly variable, and there will be long periods of time throughout which the river will be unable to move at least the coarsest part of its bed-load. The incidence of bar initiation, flow diversion and the creation of new channels (braiding) will thus be high. During high discharge events in streams with high stream power, "bank erosion is vigorous, and a wide braided channel forms" (Schumm 1977, p. 129). The conditions of abundant, coarse, non-cohesive bed-load, strongly fluctuating discharge and steep slope are not typical of any particular tectonic or climatic setting.

Parker (1976) carried out a theoretical examination of the conditions of meandering, braiding and anastomosis. He noted (pp. 476–477) that

> It has traditionally been held that braiding is caused by sediment loads so high that the river cannot carry the total amount, resulting in deposition on the bed as internal bars and general channel aggradation. On the other hand, the mechanism causing meandering is typically identified as secondary flow associated with channel curvature. If the causes of meandering and braiding were so different a unified approach would be impossible. In fact both these theories are demonstrably incorrect. The first theory implies that braided channels can never be in an equilibrium, or graded state, whereby the load supplied from upstream of a point is balanced by the load transported downstream. Presumably, then, aggradation occurs until a higher equilibrium slope is obtained, at which point braiding must stop. However, slope increases are in fact observed to exacerbate braiding rather than damp it. Furthermore, many braided rivers do not aggrade. … As regards meandering, it has been shown herein that the channel curvature needed to induce secondary flow is a result rather than a cause of initial meandering tendencies in straight channels, a fact that has been experimentally verified.

Parker (1976, p. 477) went on to note that "most streams have a tendency to form bars even though they are in a graded state. If the slope and the width-depth ratio at formative discharges are sufficiently low, meandering is favoured"; and further, "that aggradation, by increasing the slope and forcing the channel out of its banks, can lead to a transition from a meandering to a braided state, or can increase the tendency for braiding.

Friend and Sinha (1993) carried out careful measurements of sinuosity and a braiding index (of their own devising) on three river reaches in India. They averaged these readings over 10-km reaches, measuring up to 28 individual reaches for the three rivers. The data showed considerable variation in these two parameters, which they attributed to variability in the availability of bed-load sediment, reflecting local controls of tributary and bank materials. These results are consistent with the earlier conclusions of Carson (1984), who demonstrated that bed load grain-size influences the meandering-braided transition, with coarser bed load causing the transition to occur at higher slopes and/or stream power. As Schumm (1981, pp. 26–27) noted, quite minor variations in sediment load, which could reflect changing tributary or bank conditions, may cause local abrupt temporal or downstream changes in fluvial style from braided to meandering, or the reverse. In a simulation experiment, Stølum (1996) demonstrated that the sinuosity of a meandering river varies between about 2.5 and 4 as a result of continuous changes in the sinuosity of individual channels and the effects of meander cutoffs, which locally, temporarily reduce sinuosity.

An alluvial data base compiled by Coleman and Wright (1975, Table 2.1) shows that on a world-wide basis braided rivers are equally as common as meandering rivers in Arctic, temperate, dry tropical and humid tropical regions. There are two main reasons for this:

(1) The sediment load and discharge characteristics of a river may partially reflect the climate and relief of a source area many hundreds of kilometres distant; for example tropical rivers such as the Ganges, Brahmaputra and Mekong, all of which are strongly braided, have headwaters located in the Himalayan Mountains.
(2) The causes of a high bed-load and a fluctuating discharge are diverse. The following are some of the main ones:
    (a) Alpine source areas provide strong relief and a predominance of mechanical over chemical weathering for the generation of coarse clastic debris. Discharge is markedly peaked during periods of spring snow melt. Glacial outwash streams are almost invariably braided. The deposits they form are referred to as sandurs (and were the subject of several excellent studies of braided-stream sedimentation: see Miall 1977).
    (b) Marked discharge fluctuations also are characteristic of Arctic, arid and monsoonal climatic areas.
    (c) A lack of vegetation in a drainage basin means a lack of water and sediment-storage capacity, and consequent immediate response to storms in the form of flash floods. The degree of vegetation cover in an area is mainly controlled by climate, but removal of the cover by deforestation or fire can produce the same catastrophic flooding effects as characterize arid regions (Chawner 1935).
    (d) During most of geological time, until the Early or Middle Devonian, land vegetation had a very restricted distribution (Seward 1959; Davies and Gibling 2010a, b). Sediment transportation characteristics would, therefore, have been similar to those of modem arid regions—a predominance of bed-load rivers with strongly fluctuating discharges (Schumm 1968a, p. 1583).

Yalin (1992), whose discussion of the morphology of rivers adopted the formal theoretical approach of the engineer, described the meandering condition, following Leopold and Langbein (1966) as that “form in which a river does the least work in turning.” Experimental and field observation show that meanders develop by bank erosion triggered by horizontal turbulent bursts, which develop at a regular spacing depending on the scale and discharge of the channel. These bursts may or may not be associated with localized erosion and sediment transport to develop alternate bars. It is a common observation that even straight channels commonly have meandering thalwegs with alternate bars on the insides of the incipient meanders, but Yalin (1992, p. 170) made the point that alternate bars are not always present in rivers that contain meanders. It is therefore the turbulent bursts that are the key element in the generation of meanders. Yalin (1992, p. 171) described alternate bars as the “catalysts which accelerate the formation of meanders.” Braiding develops by erosion on both banks, leaving a central elevated area, which entraps sediment, forming a braid (this is in elementary terms the process described by Yalin, 1992, p. 209). This description is very similar to that developed by Leopold et al. (1964) based on their flume model of the braiding process. According to Yalin (1992) The process is initiated (or enhanced) by steepening of the valley slope or by increase in the sediment load.

In a research synthesis by Knighton (1998) it is stated (p. 220) that “there is as yet no completely satisfactory explanation of how or why meanders develop.” The author goes on to summarize Yalin’s “theory of macrotubulent flow and the bursting process”, but argues that field tests of the hypothesis are lacking. Knighton (p. 223) returns to the idea of instability between the flow and the channel boundary resulting in the formation of alternate bars that then focus erosion, leading to meander development. The conditions for braiding are described as an abundant bed load, erodible banks, a highly variable discharge and steep valley slopes (Knighton 1998, pp. 231–232). It is suggested that most or all of these conditions need to be satisfied in order for the braiding pattern to develop. Anastomosed rivers develop in areas of low gradient, small stream power, cohesive banks, and net aggradation. Knighton (1998, p. 237), citing Smith and Smith (1980) and Smith (1983), describes these as “cohesive-sediment anabranching rivers.”

Our understanding of anastomosed rivers also undergone some evolution since the 1980s. According to Makaske (2001, p. 151), Schumm (1968a) may have been the first to point out that the term “anastomosing” should not be used as a synonym for braiding: “The terms braiding and anastomosing have been used synonymously for braided river channels in this country [the US], but elsewhere, particularly in Australia, anastomosing is a common term applied to multiple-channel systems on alluvial plains. The channels transport flood waters and, because of the small sediment load moved through them, aggradation, if it is occurring, is a slow process. As a result, these low-gradient suspended-load channels are quite stable” (Schumm, 1968a, p. 1580). The Australian rivers Schumm was describing occupy a completely different geomorphic and climatic setting than the anastomosed rivers described by Smith and Smith (1980).

As is now apparent, there are at least two distinct anastomosed fluvial styles to consider, those in humid settings (e.g., Columbia and Alexandra rivers, British

Columbia; Cumberland Marshes, Saskatchewan; Magdalena River) and those in arid settings (e.g., Cooper Creek, Australia). In humid systems an assemblage of narrow, intersecting, ribbon sand bodies develops, encased in significant thicknesses of overbank deposits, including crevasse splays, flood basin muds and, possibly coals. This is the classic facies assemblage of the Cumberland Marshes (Figs. 3.4, 3.6). Dryland systems, as illustrated and described by North et al. (2007) generate quite different facies assemblages. Channels are poorly defined, and avulsion, as typically understood, does not occur. Rather, the entire floodplain may be engulfed by rare floods, spreading sediment and water across the width of the valley. It is thought that the facies distinction between channel and overbank is much less significant than in humid systems, although stratigraphic information from modern systems is sparse and data from the ancient record absent.

There is some suggestion that the anastomosed style is a temporary one. The condition of multiple channels with frequent avulsion is a response to increased sediment load or to repeated tectonic disturbance, such as subsidence events or rise in sea level that increase accommodation and lead to increased aggradation rates. The apparent stability of anastomosed channels (for example, the lack of evidence of lateral accretion) may simply be a factor of the rate at which changes in channel position by avulsion take place, before significant channel evolution has occurred. Given stable conditions, it can be shown that the anastomosed condition tends to evolve into a meandering style. This appears to have been the post-glacial history of the Rhine-Meuse system (Törnqvist 1993). Makaske (2001, p. 169) suggested that "At the moment, it may be most appropriate to characterize long-lived anastomosis in general as a state of dynamic equilibrium, with avulsions maintaining a multi-channel system, while older channels are slowly abandoned."

Makaske (2001, p. 187) argued that not all ribbon sandstones interpreted as the deposits of stable channels are the product of an anastomosed environment. North et al. (2007) developed this idea further, pointing out the difficulty in demonstrating the action of simultaneous channels from the rock record. They re-examined the Cutler Group, New Mexico, deposits interpreted by Eberth and Miall (1991) as the product of an arid anastomosed system, and argued that the evidence for anastomosis is virtually non-existent. The fact that ribbon sandstones have been mapped intersecting each other does not demonstrate that they were contemporaneous. The Cutler facies assemblage of distinctly contrasting channel and overbank deposits, with well-defined channel "wings" and levees, is inconsistent with the details of such arid-system rivers as Cooper Creek, as documented by North et al. (2007). Ribbon sandstones are the predominant architectural form for channel sandstones in the Cutler Group, but these probably are the product of a single-thread fluvial system, such as occur in some large, sandy alluvial fan settings.

One of the most recent treatises on fluvial sedimentology is that by Bridge (2003), in which much of the evolution of ideas (including that discussed here) about the causes of meandering and braiding, and the controls on fluvial style, is challenged. Bridge (2003, pp. 153–154) rightly notes the importance of the "channel-forming discharge" in determining fluvial styles. This category of event may correspond to the bankfull discharge or to the seasonal maximum discharge, or to some rarer, larger-scale event, depending on the river system. Bridge (2003, pp. 154–155) argues that "the degree of braiding and the width/depth of channels increase as water discharge is increased for a given slope and bed-sediment size, or as slope is increased for a given water discharge and bed-sediment size", a statement which encapsulates the conclusions of the earlier generation of geomorphologists (Leopold and Wolman 1957; Leopold et al. 1964). However, Bridge then goes on to categorize as myths several long-held conclusions regarding fluvial style:

> A common myth is that discharge variability is greater for braided rivers than for single-channel rivers. This myth probably originated from the early studies of proglacial braided rivers in mountainous regions of North America, where discharge varied tremendously during snowmelt. In contrast, many single-channel rivers were studied in temperate lowland regions where discharge variations were moderated by groundwater supply. It is very clear, however, that discharge variability does not have a major influence on the existence of different channel patterns, because they can all be formed in laboratory channels at constant discharge, and many rivers with a given discharge regime show along-stream variations in channel pattern (Bridge 2003, p. 156).

This statement seriously oversimplifies the earlier examinations of the causes of braiding. Discharge variability, as noted in earlier paragraphs, is only one of several key variables that cause braiding. The nature of bed and bank materials, the influence of merging tributaries, and the bedrock control of valley slope, are all contributing factors; so, yes, channel style may vary along a river that is characterized by the same discharge pattern throughout, because, for example, the bank stability varies with varying sediment composition.

A little further on, Bridge (2003, p. 157) claims that "the correlation between sediment load, and bank stability is not generally supported by data, as recognized implicitly by Schumm (1981, 1985) in his more recent classifications of channel patterns." However, Schumm's classification, which is reproduced here (as Fig. 2.18) from Knighton's book, where it is described favorably as encompassing much of the full range of natural variability, makes it very clear that stability, sediment load and sediment size are all important influences on fluvial style. It is the combination of variables that controls channel style, not any single parameter in isolation. This is one reason why interpretations of channel style in the rock record are so difficult to make and so difficult to interpret.

Jerolmack and Mohrig (2007) employed two relationships to generate a descriptive discrimination between the main fluvial styles (Fig. 2.19). They turned to Parker's (1976) stability criterion to assess whether a river should be braided or meandering (single-channel with multiple high-velocity threads, or single-thread with straight to sinuous planform). This criterion, ε is defined as $\varepsilon = S\sqrt{(ghB^4)}/Q$, where S = water slope, g = gravitational acceleration, h = channel depth, B = channel width, Q = formative water discharge. The Mobility number, M, is defined as the ratio of avulsion and lateral-migration time scales $M = T_A/T_C$. Rivers consisting of a single channel that sweep across a flood plain, reworking the floodplain by lateral erosion and deposition have a value of $M \gg 1$. In these cases, avulsion is infrequent. Where rivers are undergoing active aggradation, with frequent avulsion, several channels may be active at once. Such systems have values of $M \ll 1$.

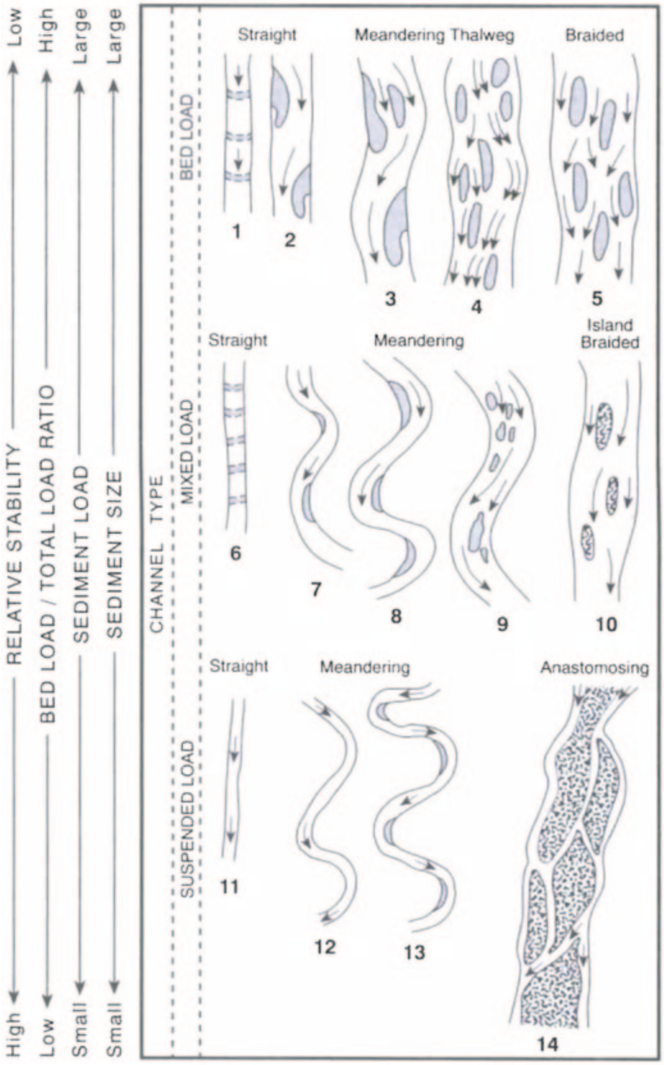


Fig. 2.18 A classification of channel patterns, adapted from Schumm (1981, 1985)

The various discussions of the controls on fluvial style summarized here (a discussion which is by no means exhaustive) do not provide much in the way of useful ideas for geologists to follow. Geomorphic controls, such as valley and channel slope, discharge and sediment load, are not amenable to exploration and documentation for the geological record, so it is not much use for reservoir geologists to wrestle with the implications of the structure of turbulent bursts in the formation of meanders (I realize this sort of statement is anathema to the sedimentological purist, but I am writing here for the practicing petroleum geologist who want tools

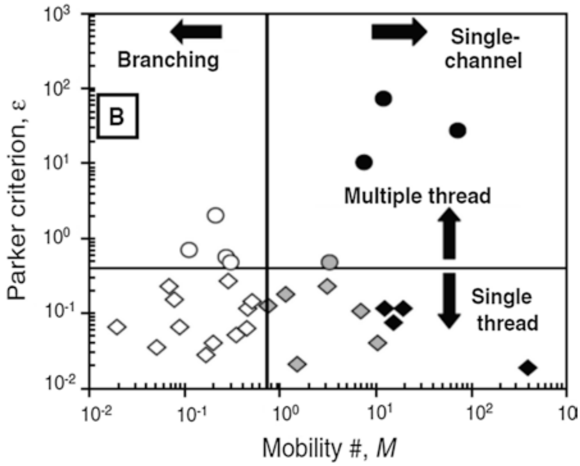


Fig. 2.19 Plot of Parker's (1976) stability criterion against Mobility number for thirty modern net-depositional systems identified by Jerolmack and Mohrig (2007). River styles: *diamonds* = sinuous, *single-thread*; *circles* = braided; *white circles* = single channel, *black circles* = branching, *gray circles* = transitional (Jerolmack and Mohrig 2007, Fig. 2b, p. 464)

that are useful). Perhaps the whole endeavour has been approached from the wrong end. While we are conditioned by the Hutton-Playfair-Lyell tradition to cite that mantra "The present is the key to the past," uniformitarianism needs to include the caveat that the past is a very selective record of the present, and that this selectivity is determined by geological factors that cannot be directly observed at the present because they involve the additional factor of geological time and the issue of preservability (see Miall, in press, for a more general discussion of this topic).

What we really need to do is to look at what has actually been preserved and to see if we can understand what to make of it. Here we enter a rich literature written by practical sedimentologists who have looked long and hard at the actual ancient record. It is, after all, what gets into the record, not the ephemera of the present day, that is important to reservoir geologists. It was Geehan (1993, p. 56), a practicing petroleum geologist, who stated "Clearly, outcrops are the only source of geological analogue data that show indisputably what is preserved in the geological record, in a form that fully represents all scales of heterogeneity up to the size of the outcrops. Thus, outcrop data must continue to provide our most reliable controls for modeling aspects of reservoir heterogeneity that are not directly measured in the subsurface."

2.3.4 Architectural Classifications Based on the Ancient Record

Classifications that focus on empirical description and two-or three-dimensional information about preserved fluvial deposits require excellent outcrop, and it is therefore not surprising that the first attempts at such classifications emerged from research being carried out in the Cenozoic foreland-basin deposits flanking the Pyrenees in northern Spain, where a relatively arid climate and limited vegetation cover have created excellent-large-scale outcrops much studied by Spanish, Dutch and British geologists.

One of the first of these attempts to focus on the empirical record of ancient fluvial deposits, as a basis for classification and interpretation, was that by Friend et al. (1979). They pointed out the inadequacies of such style-related terms as "braided" and "meandering" and suggested three ways of examining and classifying ancient fluvial deposits, of which fluvial style is but the first. The second category of description refers to the shape of the preserved sandstone or conglomerate bodies; these may be narrow, channelized units, commonly ribbon-shaped, or broader, sheet-like units (Fig. 2.20). Friend et al. (1979) suggested a distinction be made between sheets and ribbons at a cut-off value of 15 for the width to thickness ratio. Their third category of description related to the internal architecture of the sandstone or conglomerate body, making note of whether the unit consists of a single stacked succession or whether it is complex, comprising several or many individual successions or "storys" bounded by internal bounding surfaces (Fig. 2.20). This last category of description touches on the method of architectural-element classification that evolved from the work of Allen (1983), and is discussed in the next section.

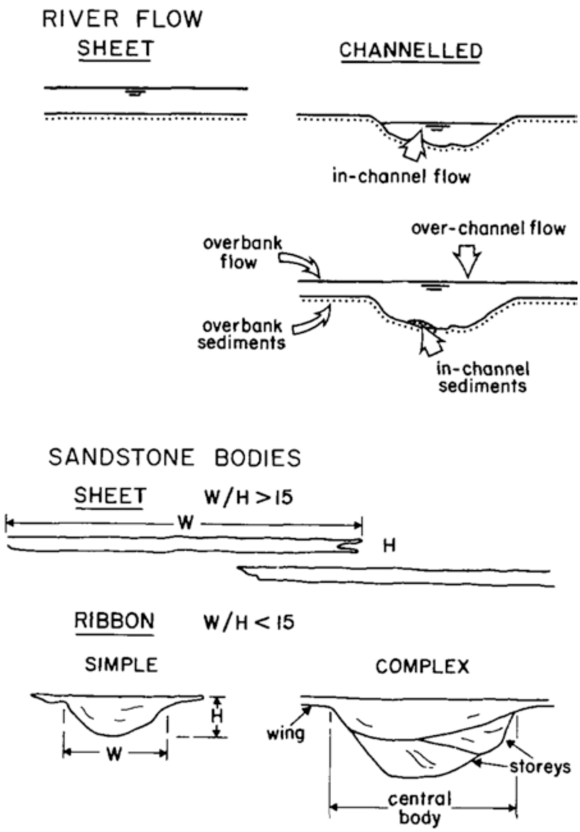


Fig. 2.20 The classification terminology suggested by Friend et al. (1979) for the description of mid-Cenozoic sandstone bodies in the Ebro Basin, Spain

Some elaborations of this preliminary approach to classification were suggested by Friend (1983). He proposed using the non-genetic term "hollow" rather than channel (Fig. 2.21), and added the term "mobile-belt" to the terms ribbon and sheet, for the external geometry of sandstone bodies. Ribbon bodies were assumed to represent fixed channels, while mobile channel belts implied the lateral movement of the channel (or channels), resulting in the lateral amalgamation of channel-fill units. He suggested that mobile channel belts may develop by steady lateral migration or by migration and channel switching (Fig. 2.22). As would now be recognized, much also depends on the rate of basin subsidence. Rapid rates of subsidence might result in the development and preservation of floodplain deposits contemporaneous with channel migration, resulting in a separation of channel units into separate bodies, whereas slow subsidence might lead to a channel combing back and forth through its own deposits and developing a broad, sheet-like unit.

Friend's final classification scheme (Fig. 2.23) uses terms such as braided and meandering, but only for description of channel behavior, not as terms for the description of the alluvial architecture. He followed Schumm (1963, 1968a) in recognizing the importance of the nature of the sediment load and the character of bank materials as controls on the resulting architecture.

The application of Friend's approach to the Pyrenean outcrops lead to the sandstone classification of Hirst (1991), reproduced here as Fig. 2.24. Note the

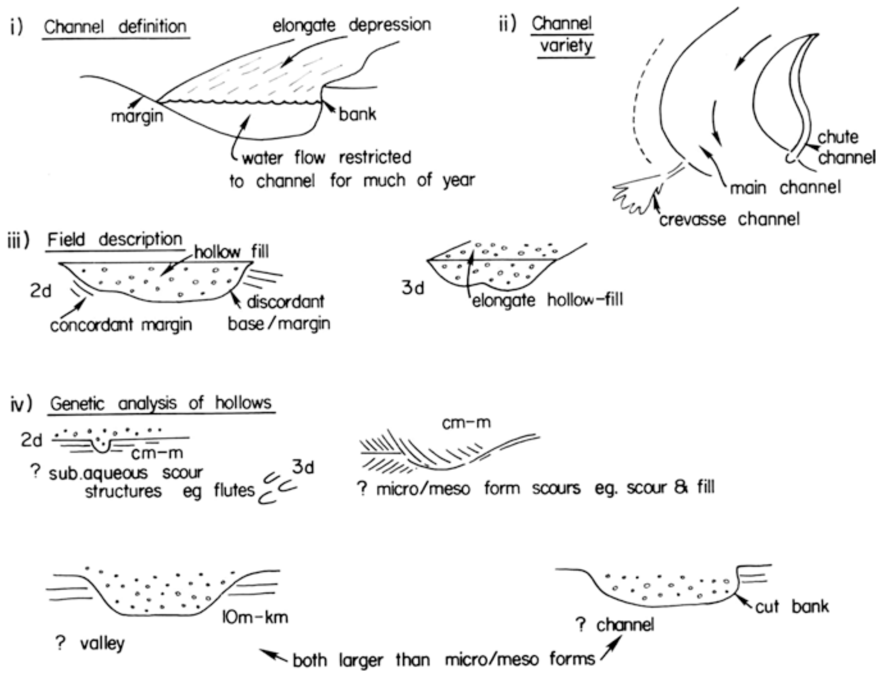


Fig. 2.21 Definition of "channel" and related terms (Friend 1983)

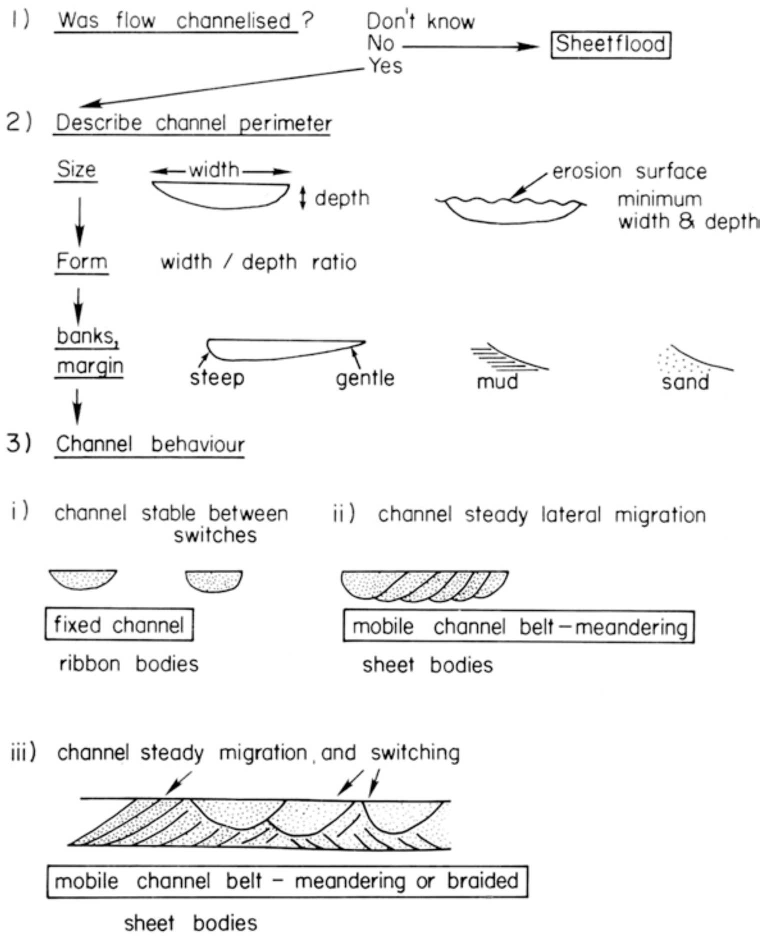


Fig. 2.22 Classification of sand bodies, and terms used (Friend 1983)

introduction of the term "wings" for thin wedge-shaped sandstone units that extend laterally from channel margins. These are interpreted as levee deposits. Hirst (1991) used this classification in his studies of the laterally very extensive outcrops of Cenozoic fluvial systems in northern Spain.

Galloway (1981) developed his own approach to classification of alluvial architecture, based on his studies of Cenozoic successions along the coast of the Gulf of Mexico. These have been studied largely in the subsurface and, as Galloway (1981, p. 133) stated:

> For regional synthesis and description of complex fluvial sequences, classification by type models (such as braided, fine-grained meander belt, etc.) proves to be of limited use. First, the variety of channel-fill deposits encountered typically includes many examples that show little resemblance to well-described modern analogs. Secondly, a variety of

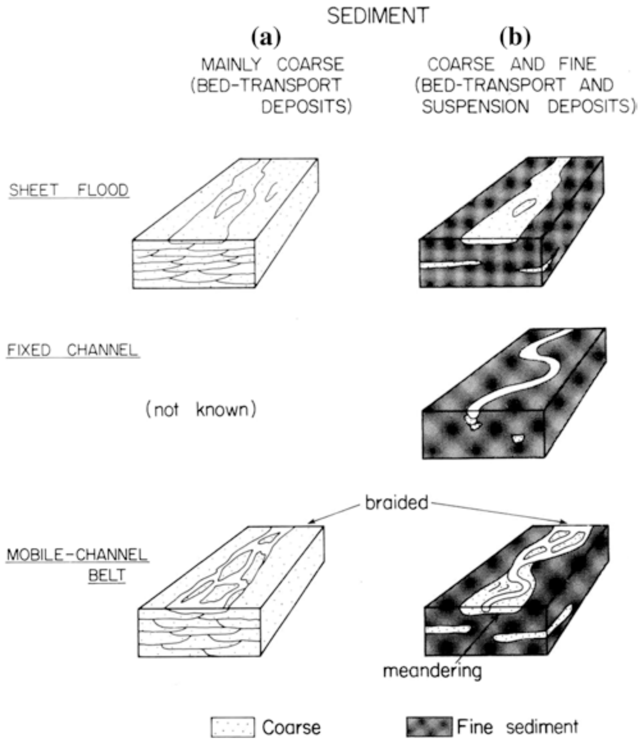


Fig. 2.23 Friend's (1983, Fig. 6) classification of alluvial architecture

> intergradational fluvial sequences typically occurs at the same stratigraphic horizon within a given local area. … Thirdly, use of models requires detailed description of vertical or lateral textural and sedimentary structural sequences that cannot be obtained in the subsurface, or in areas of poor exposure.

Galloway (1981) based his approach on the classification scheme of Schumm (1977, 1981), which emphasizes the importance of the type of sediment load transported by the rivers. This system focuses on the subsurface characteristics of the various preserved styles, particularly sand isolith patterns, vertical profiles and predicted lateral relationships (Fig. 2.25). Architectures comparable to the ribbon, multistory and sheet categories of Friend and co-workers, are apparent in this classification.

Another practical approach to classification of alluvial architecture was offered by Alexander (1993), based on the surface and subsurface study of Jurassic units in Yorkshire, U.K., and the North Sea Basin (Fig. 2.26). Again, the classification is built on that of Friend et al. (1979) and Friend (1983). She wrote (p. 152):

> No natural system is (or was) steady state; fluctuations in river discharge, temperature, tidal variations, and storm surge are some of the major factors that result in major shifts in facies belts on a variety of time scales. These changes are more extreme where they are superimposed on a progressive change resulting from eustatic, tectonic or other prolonged relative sea-level change.

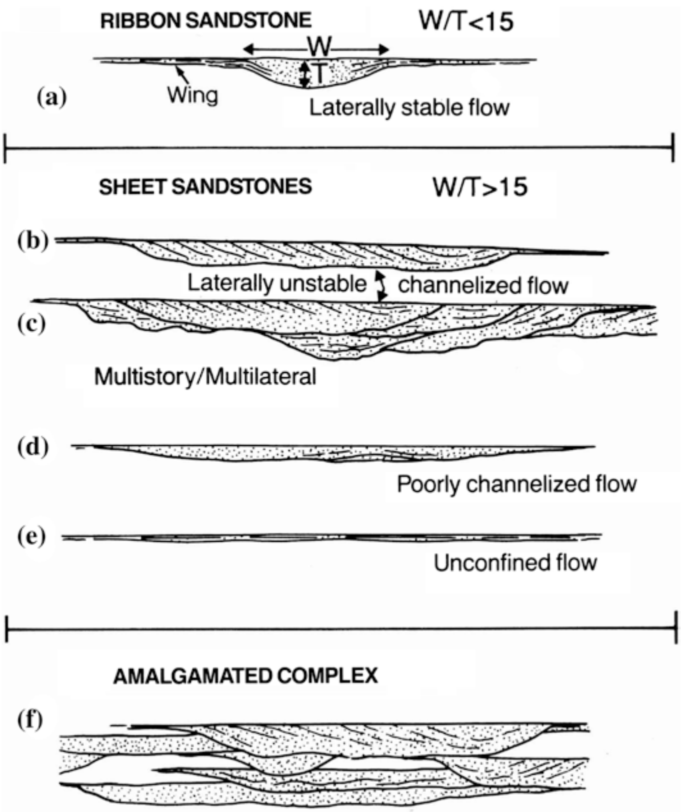


Fig. 2.24 The range of sandstone-body geometries in the Huesca fluvial system (Oligocene-Miocene), Ebro Basin, Spain (Hirst 1991)

As she noted (p. 155): "It is unwise to infer the channel plan form from limited two dimensional exposure." This classification is strictly empirical, and ideally suited to a methodical, analytical approach and to the production of architectural documentation free of dogma that may then be used for production purposes.

Gibling (2006, p. 731) argued that although there had been many studies of the internal organization of channel deposits, "in contrast, only a few accounts … have dealt comprehensively with the dimensions and 3-D form—or *external* geometry—of channel deposits and valley fills." He noted the importance of this topic to students of sequence stratigraphy, as well as the traditional needs of the resource-industry explorationist. Gibling (2006) assembled a data base of more than 1,500 examples of channel-fill deposits, ranging in age from Precambrian to Quaternary. These constituted a highly variable data set, including channel-belt units and valley-fill deposits deposited under a range of low- and high-accommodation settings. He made use of Potter's (1967) terms *multistory* and *multilateral* (Fig. 2.27) for units developed by vertical stacking and lateral amalgamation, respectively. *Succession-dominated*

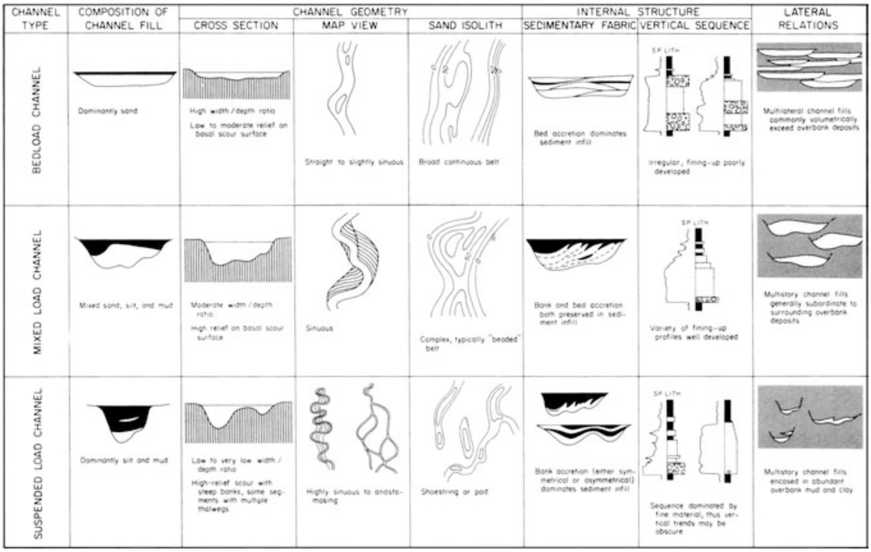


Fig. 2.25 The sand body classification scheme of Galloway (1981), which is based on the sediment-load classification of Schumm (1977, 1981)

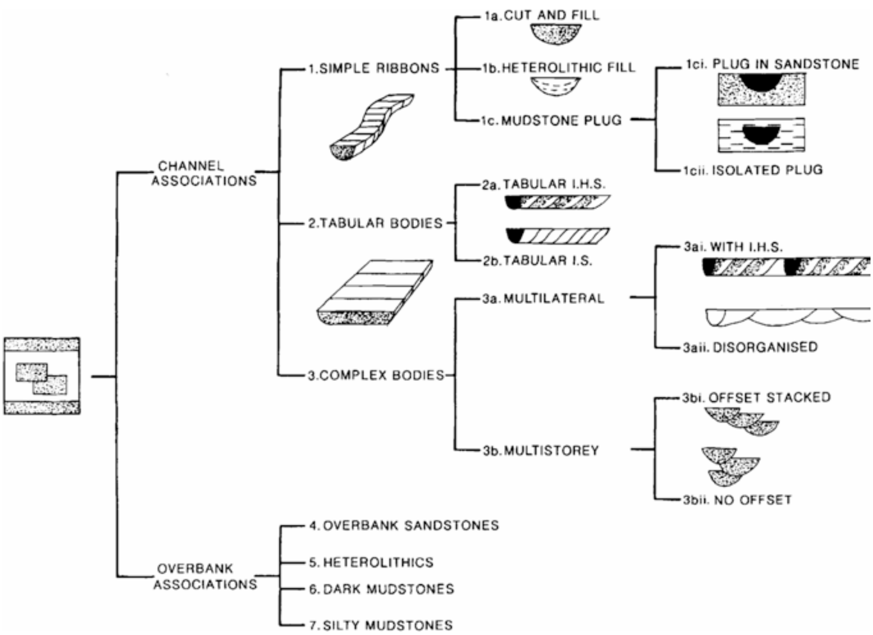


Fig. 2.26 Facies association classification diagram of Alexander (1992). The channel association architecture is independent of size. The facies associations are end members in a continuum of possibilities. Inclined homolithic stratification (I.S.) and inclined heterolithic stratification (I.H.S.) follow the usage of Thomas et al. (1987)

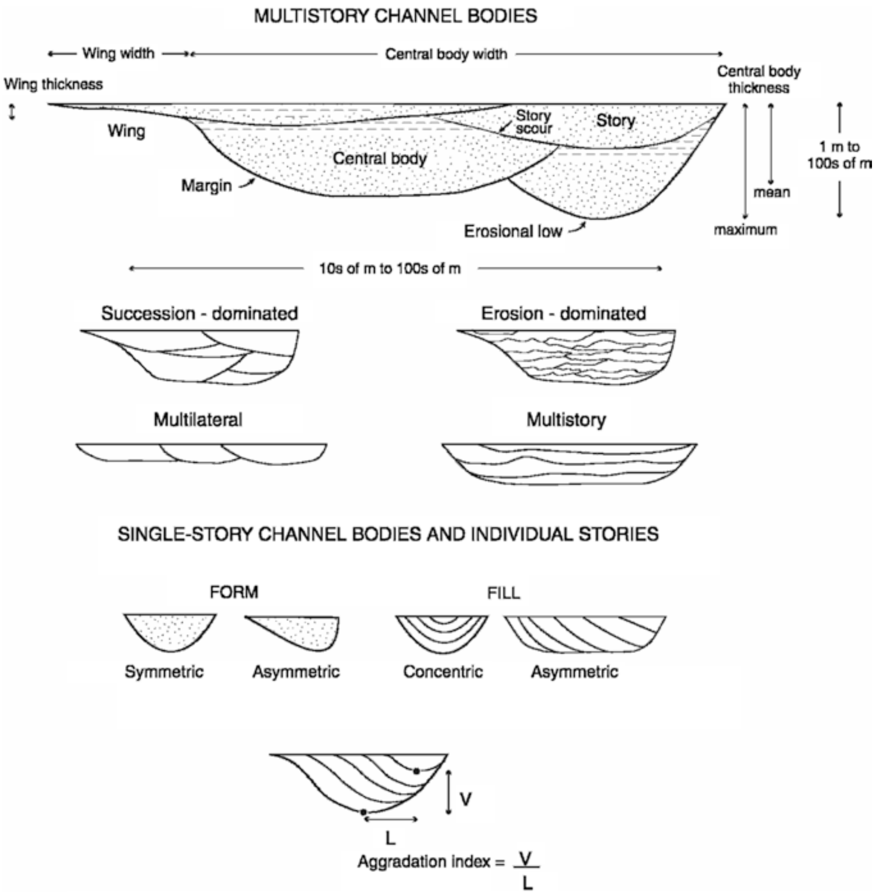


Fig. 2.27 Terminology for describing the cross-sectional geometry of channel bodies (Gibling 2006, Fig. 2)

and *erosion-dominated* are also terms used for multistory units which highlight a distinction between bodies preserving relatively complete channel-fills versus those comprising several or many fragmentary deposits separated by erosional bounding surfaces (Fig. 2.27). The details of channel architecture may incorporate many geomorphic elements that are difficult to recognize from the ancient record. Amongst the most important and least appreciated are scour surfaces. As Best and Ashworth (1997) noted, scour depth at channel confluences and at bends may be as much as five times greater than mean channel depth. Scour fills have been recognized as forming a distinct type of architectural element—the scour hollow (element HO of Miall, 1996, 2010a), following the work of Cowan (1991). Width and thickness of channels may vary dramatically in short distances as the channel enters bends, encounters tributaries, or resistant bank materials. W/D data may be affected accordingly. The complete classification is shown in Fig. 2.28.

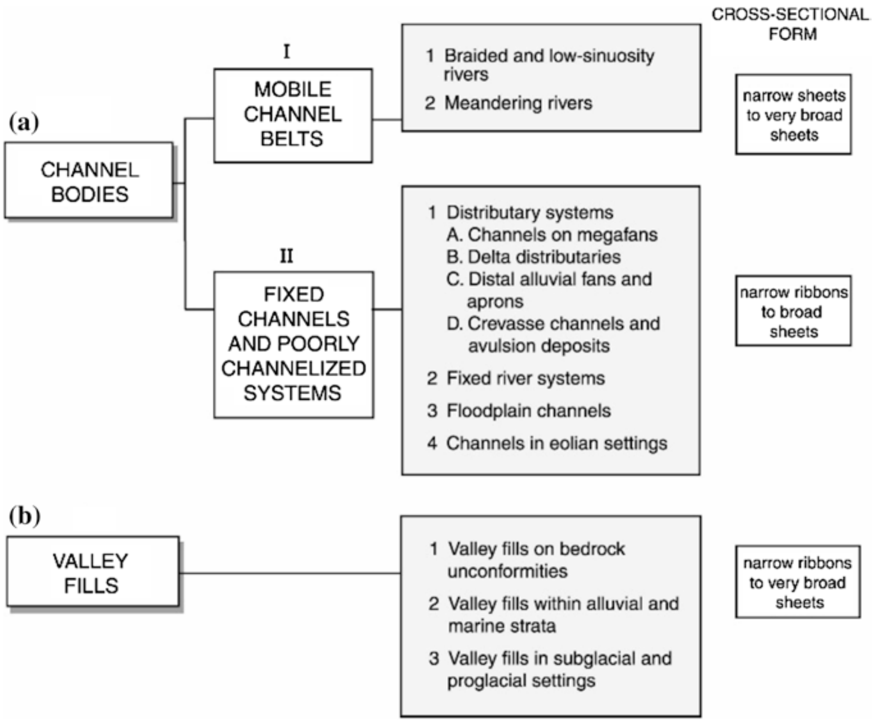


Fig. 2.28 Classification of fluvial channel bodies and valley fills based on dimensions, geomorphic setting, and architecture (Gibling 2006, Fig. 4)

Gibling's data indicate that the range of widths and thicknesses for braided and other low-sinuosity systems, and those of meandering systems, overlap extensively at the lower end of the width-thickness envelope (Fig. 2.29). Widths range from 30 to 10,000 m and thicknesses from 2 to less than 100 m, except for some outliers (some wider and thicker braided systems). The average width/thickness ratio for meandering systems is in the range of 1:100; for braided systems about double that. As noted by Gibling (2006, p. 737), "the larger channel bodies are mainly those of meandering and braided rivers, which tend to generate wide sheets." Gibling (2006, p. 753) noted, interestingly, that "meandering rivers do not appear to create thick or extensive deposits and, despite their familiarity in modern landscapes, their deposits probably constitute a relatively minor proportion of the fluvial-channel record." This is worth emphasizing, given the prominence of the familiar meandering river channel and its point-bar complex in many sedimentology textbooks—for example, this is the illustration that has been used for the cover of the Geological Association of Canada's "*Facies Models*" volume, in its various editions, since it was first published in 1979. It is noteworthy, in this regard, that Blum et al. (2013) have reached the opposite conclusion:

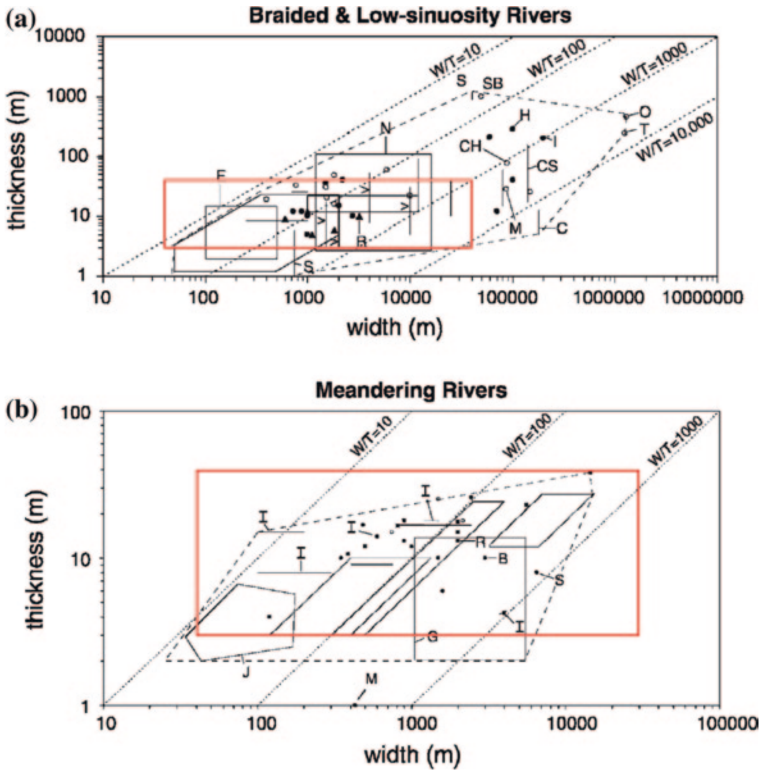


Fig. 2.29 The range of widths and depths of two major classes of preserved channel sand body, as compiled by Gibling (2006, Fig. 6). The *red rectangles* enclose the same size range for each of the two plots: thickness range: 3–40 m, width range: 40 m–30 km

> Experiments show that braided channels are the default self-organizing pattern in non-cohesive sediments, and that self-sustaining single-channel meandering patterns require bank-stabilizing muds and/or vegetation that reduce rates of outer bank erosion … However, on a global scale, braided channels are relatively uncommon (Paola et al. 2009), the majority of channels in modern subsiding basins are meandering or anabranching, and anabranching patterns dominate large low-gradient river systems.

These diverging opinions may be a reflection of the different experiences of the two authors. Blum has worked primarily with modern and post-glacial systems, whereas Gibling's experience has encompassed a wide range of modern and ancient systems. Both note that the meandering patterns is common and familiar in the modern landscape, but whereas Blum seems to assume that this prominence would be reflected in the rock record, Gibling, having studied numerous ancient systems for his 2006 compilation, suggests otherwise.

Gibling's (2006) study contains useful comments regarding the recognition of valley-fill deposits, which offer particular challenges to the subsurface stratigrapher. Bounded as they are by erosion surfaces, just like any channel fill unit, the recognition of the particular origin of valley-fills may be difficult. Gibling (2006, p. 742) suggested three diagnostic criteria for valley-fills: (1) the presence of a widespread basal erosion surface; (2) the dimension of the valley-fill are an order of magnitude larger than any component or otherwise correlatable channel deposit; (3) the scale of erosional relief at the base is several times that of the typical channel fill. As Posamentier (2001) and Miall (2002) demonstrated, incised valleys may be flanked by incised tributaries and gullies (e.g., Fig. 4.43). The development of valley-fills may be a consequence of base-level change or climate change, and their analysis then becomes an integral part of the study of sequence stratigraphy (Chap. 6).

Gibling (2006, p. 760) noted the importance, in general, of allogenic controls on alluvial architecture. For example, "For single-story channel bodies with a given initial aspect ratio, the balance between bank migration rate and channel aggradation rate determines to a first approximation the channel-body geometry." Aggradation rate, which depends on base-level change and sediment supply considerations, is also a major control on avulsion, and therefore on the rate of initiation of new channels. Drawing on the results of this study, and the results of modeling experiments (e.g., Paola 2000), Gibling (2006, p. 763) concluded that these observations "tend to suggest that fluvial channel bodies in the geological record represent a geomorphic spectrum and that alluvial basin-fill stratigraphy is largely controlled by these factors and not by channel morphology." The importance of this conclusion cannot be over-emphasized. Since sedimentologists first began relying on the process-response model they (we!) have been obsessed with surface form of depositional systems, notably the familiar shapes and textures of modern rivers as seen from above. In my first paper on architectural-element analysis (Miall 1985, p. 266) I noted "It is the plan view of … macroform elements that generates the familiar fluvial channel styles, so commonly illustrated by low-level aerial photographs of modern rivers." However, it has now been demonstrated that these geomorphic features are of secondary importance and must be supplemented by considerations of allogenic processes, using sequence-stratigraphic concepts, for a full understanding of alluvial architecture.

2.4 Architectural Element Analysis

Difficulties with the standard facies models began with braided systems which, by their nature, tend to be complex and less predictable than the standard "fining-upward" point-bar successions of the basic meandering-stream model. Allen (1983, p. 237) pointed out that "Channel behaviour and type can practically never be predicted unambiguously from vertical sequences, if only because each kind of stream is capable of generating a wide variety of local sedimentological patterns." He argued that increasing attention was being paid to the shape and internal architecture of channelized units. In a study of the sandy-braided deposits of the Devonian Brownstones Formation in the English-Welsh border area, he noted the predominance of complex, interbedded, lenticular units clearly representing the deposits of various types of in-channel bar and minor channel, and this led to the recognition of "eight kinds of depositional features" or "internal architectural elements" (Fig. 2.30).

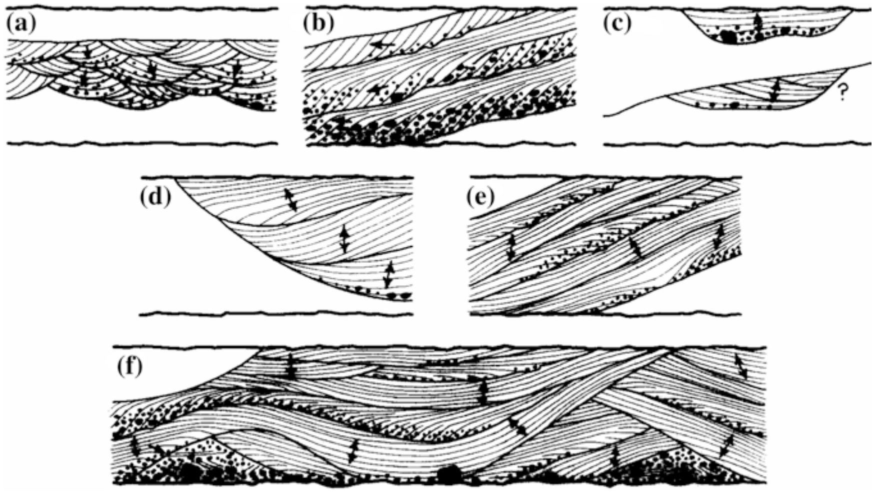


Fig. 2.30 Summary of the main kinds of depositional feature recorded from sheet sandstones preserved in the Brownstones. **a** Tabular layers of dune cross-bedded (trough cross-bedded) sandstone; **b** assemblagesof down-climbing (forward-accreting) bar units; **c** minor channel forms and fills; **d** major channelform and fill; **e** groups of laterally-accreted bar units; **f** symmetrical complexes (sand shoals) of laterally accreted bar units with gravel cores (Allen 1983, Fig. 18)

In a similar detailed study performed on large outcrops where bedding units could be traced laterally, Ramos and Sopeña (1983) and Ramos et al. (1986) defined eleven types of gravel and eleven types of sand body in a Permo-Triassic unit in Spain (Figs. 2.31, 2.32).

Meanwhile, a quite different approach to fluvial facies studies was being undertaken by geomorphologists Gary J. Brierley and Edward J. Hickin, studying the deposits of the modern Squamish River, north of Vancouver, British Columbia (Brierley 1989, 1991a, b; Brierley and Hickin 1991). This river varies along its length between braided, wandering and meandering styles. It was the purpose of their study to report the results of an intensive test of the supposed link between river planform and fluvial sedimentology in modern rivers in one particular field setting. They explained their field procedure, which was based on trenching of the modern deposits, as follows: "Four morpho-stratigraphic units are identified: bar platform, chute channel, ridge and remnant floodplain. When analysed in trenches and bank exposures, these preserved floodplain depositional units are termed elements, and the remnant floodplain unit, composed of deposits laid down by unconfined flows on bar/island surfaces, is differentiated into three top-stratum elements, namely proximal, distal and sand-wedge top-stratum elements" (Brierley and Hickin 1991, p. 74). They found that facies studies on their own were inadequate to characterize the planform style of the river, and the reverse, that planform was no predictor of facies. Element composition of the Squamish River floodplain varies from site to site, related specifically to the character and extent of sediment

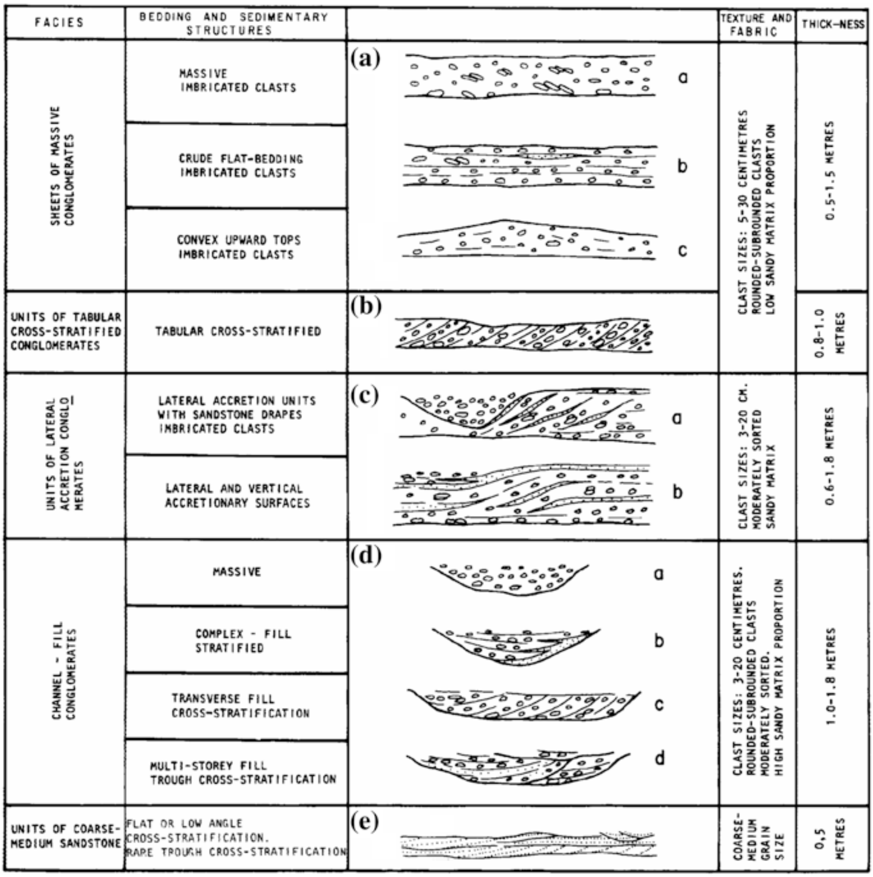


Fig. 2.31 The major types of gravel-dominated depositional feature in the Triassic Bundsandstein of central Spain (Ramos et al. 1986)

reworking and *not* necessarily to channel planform type. Amongst their conclusions is this interesting observation (Brierley and Hickin 1991, p. 81):

> Given this situation, it may be more appropriate to change the question posed above. Rather than focussing attention on planform type, it may be more appropriate to focus interpretation on mechanisms of floodplain development, examining exposures at the scale of those processes by which sediments become preserved in the floodplain.

In other words, focus on the sediments, not the fluvial style. They began to formulate what they called a "constructivist" framework, which views fluvial deposits as particular associations of elemental units, which mayor may not relate to pre-existing models (Brierley and Hickin 1991, p. 81).

These studies contained the basis of a new architectural approach, which Miall (1985) proposed could be applied to all fluvial deposits. Architectural-element

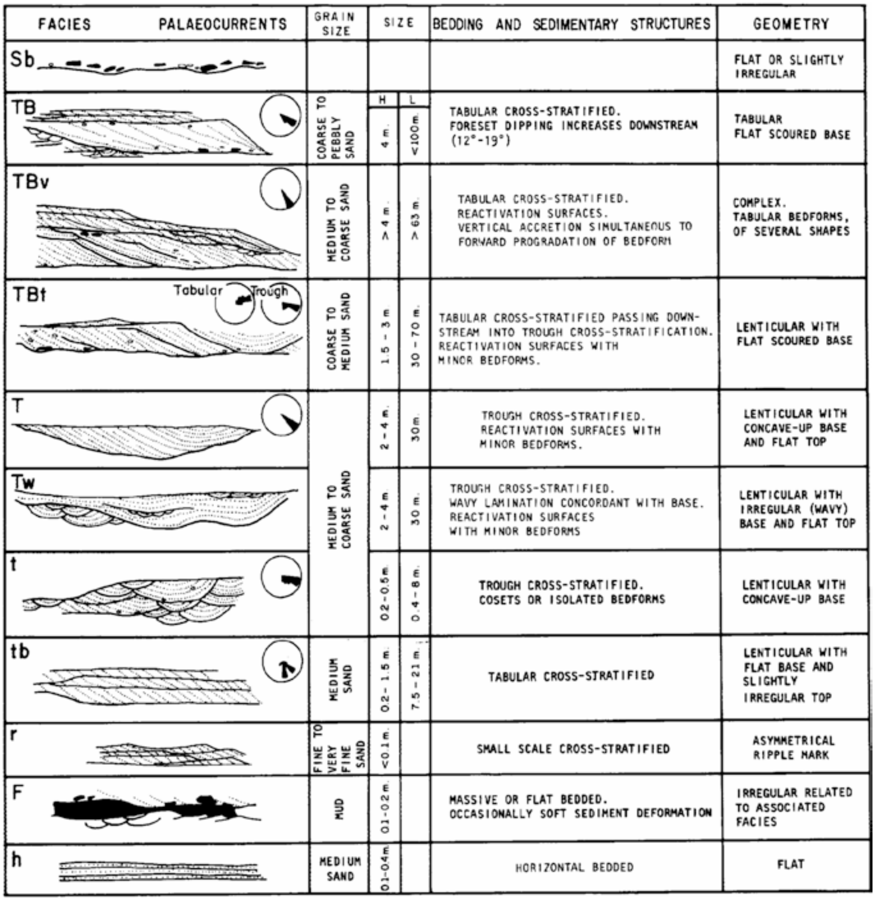


Fig. 2.32 The major types of sandstone-dominated depositional feature in the Triassic Bundsandstein of central Spain (Ramos et al. 1986)

analysis focuses on "macroforms," to use Jackson's (1975) term. These are the component units of channels and floodplains, comprising the various building-blocks of channelized and non-channelized sandstones and conglomerates, and floodplain complexes. They reflect the cumulative effect of many dynamic events over periods of tens to thousands of years. They include major and minor channels and the larger, compound bar forms such as point bars, side bars, sand flats and islands, plus such floodplain elements as crevasse channels, levees and splays.

Brierley's parallel constructivist approach is described in a book chapter (Brierely 1996). He noted (p. 276):

> Given the lack of geomorphological distinctiveness of individual channel plan form styles, it is scarcely surprising that planform-sediment correlates are far from unequivocal. Similar bedform-scale facies assemblages may be viewed independently of planform style

> and there appear to be no sedimentary structures that are peculiar to individual channel plan form types (Bridge 1985; Brierley 1989). The principle of convergence, in which depositional units stack in a similar manner for different planform styles, has also been demonstrated for element assemblages that comprise the floodplains of contiguous braided, wandering and meandering reaches of a gravel-bed river (Brierley and Hickin 1991).

Brierley (1996, p. 279) cited Collinson (1978, p. 579) to the effect that description and interpretation should not have as their primary objective to generate comparisons with existing models, but rather to construct interpretations based on the observed assemblage of elements. That this is a necessary approach was illustrated by his table (adapted here as Fig. 2.33) demonstrating that most observed element types occur in more than one planform setting, so that it is the structure of the individual elements and their overall assemblage that becomes critical in understanding the architecture of an individual deposit.

Walker (1990, p. 779) had complained that architectural-element analysis "offers no overall point of reference (*norm*) for a depositional system as a whole. Each combination of architectural elements (each individual example) is treated as unique, and in the absence of a norm, there is no way of knowing whether the individual example is similar to, or greatly different from, other examples. This is sedimentological anarchy." In fact, this is sedimentological reality, as increasing numbers of studies of fluvial deposits have made clear. This ever-expanding body of research led to the definition of sixteen facies models by Miall (1996, Chap. 8), and it was pointed out that there may be gradation and intermediate forms between any two of the models.

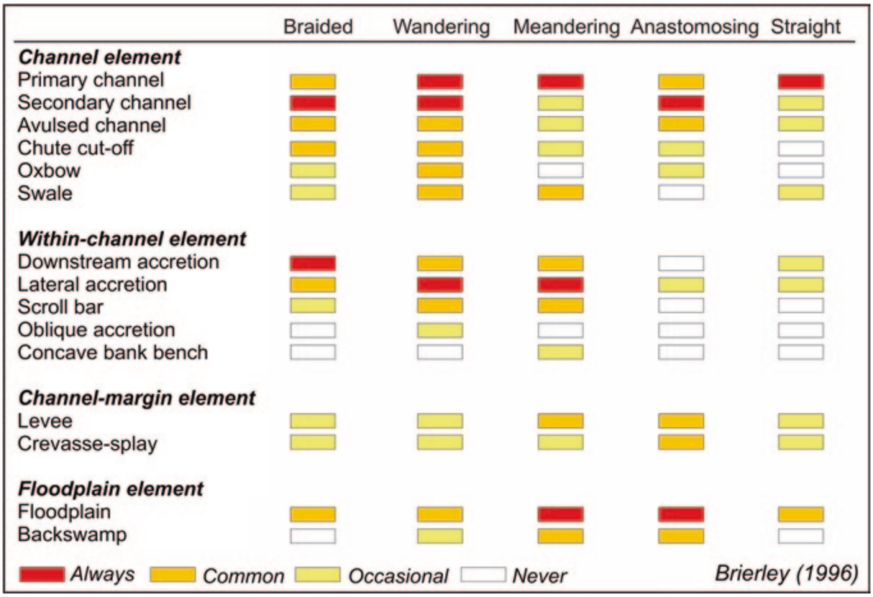


Fig. 2.33 Element presence as components of four major planform styles (adapted from Brierley, Table 8.4)

A detailed discussion of the methods of architectural-element analysis is set out in Chaps. 3 and 4 of Miall (1996), and a detailed documentation of the major types of architectural element is provided in Chaps. 6 and 7 of that book. Miall (1985, 1996) had proposed that there are eight basic architectural elements in fluvial deposits. To this has been added the hollow element (Fig. 2.34). Additional details of the lateral accretion element were provided by Miall (1985, 1996) (Fig. 2.35) and later work (Miall 1996, Fig. 7.3) expanded the "floodplain fines" element to encompass the actual variability encountered in floodplain assemblages (Fig. 2.36). A thorough documentation of architectural elements of within-channel and over-bank elements was provided in the earlier book (Miall 1996). Since that time the

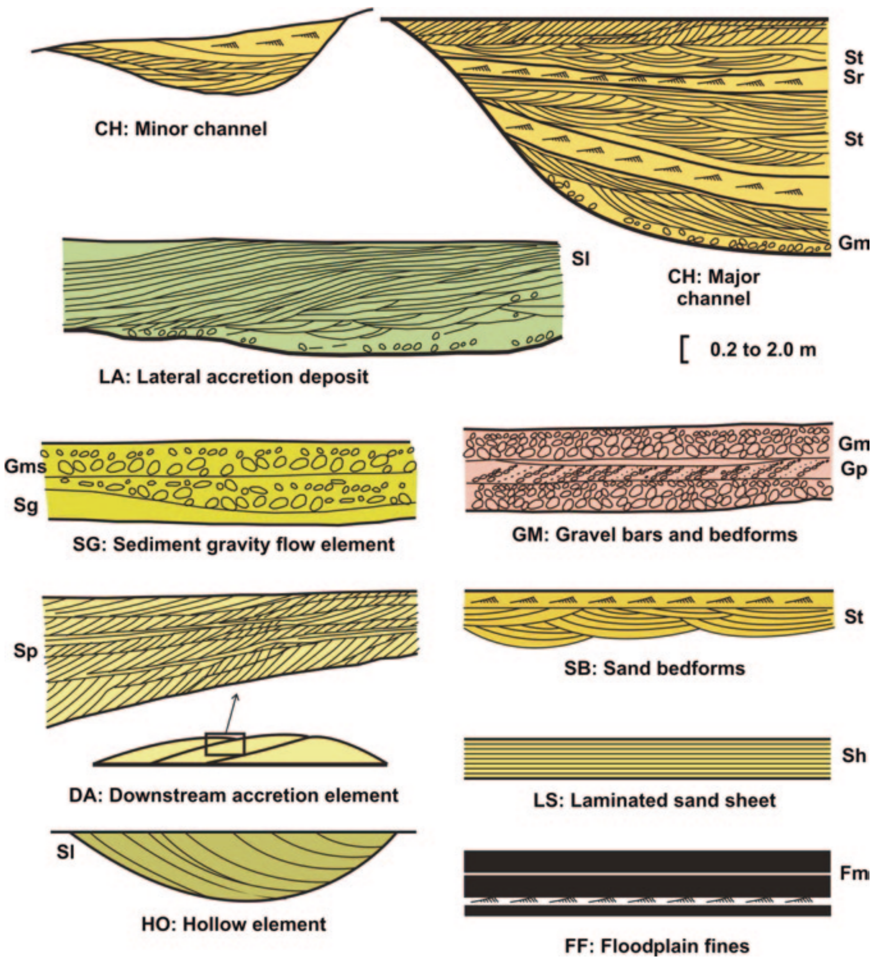


Fig. 2.34 The original eight architectural elements of Miall (1985) to which has been added the hollow element: HO

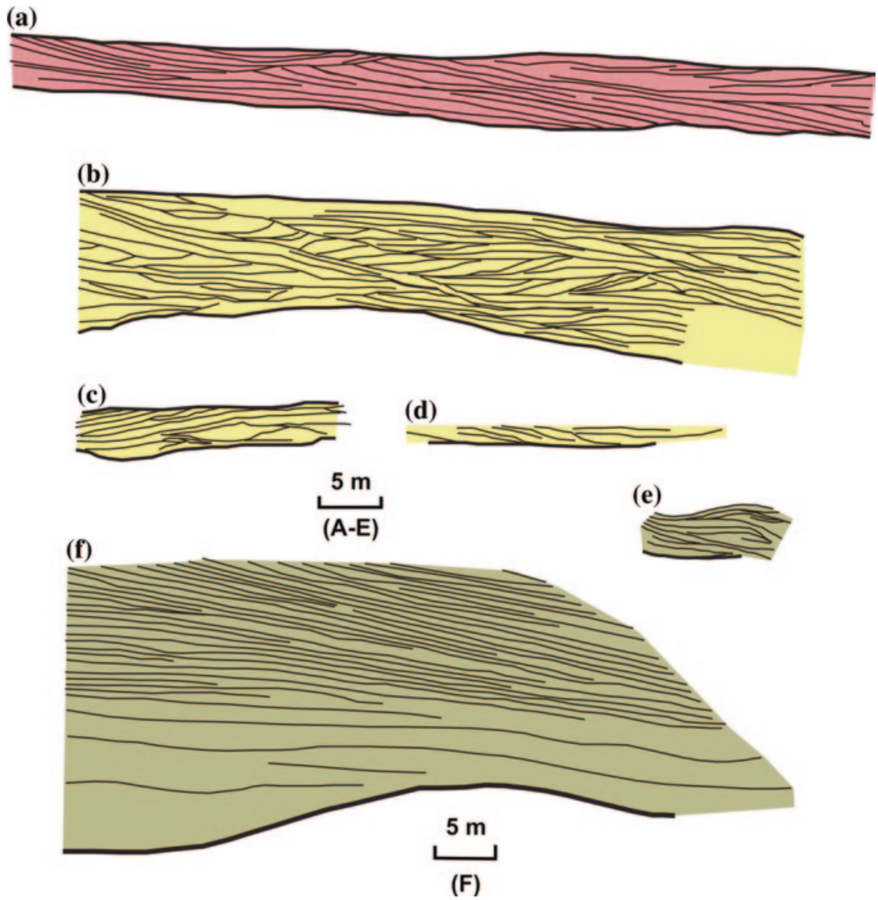


Fig. 2.35 Examples of lateral accretion elements. No vertical exaggeration. Fluvial-style model numbers of Miall (1985) are indicated. **a** Conglomerate point bar (lithofacies Gm), with chute channels (lithofacies Gt), model 4 (Ori 1979); **b** element composed of medium-grained sandstone, with abundant internal planar-tabular crossbedding (lithofacies Sp), model 6 (Beutner et al. 1967); **c** fine- to very-coarse sandstone and pebbly sandstone with cobble to boulder conglomerate lag. Abundant internal crossbedding (lithofacies Sp, St, Sh, and Sl), model 5 (Allen 1983); **d** small sandy point bar with abundant dune and ripple crossbedding (lithofacies St, Sr), model 6 (Puigdefabregas 1973); **e** point bar composed mainly of fine sandstone and siltstone (lithofacies Sl) with minor medium- to coarse-grained, crossbedded sandstone (lithofacies St) at base, model 7 (Nanson 1980); **f** giant point bar with thick, fine-grained trough crossbedded sandstone at base (lithofacies St) passing up into accretionary sets of alternating fine sandstone and argillaceous siltstone showing evidence of tidal bundling (lithofacies Se), model 7 (Mossop and Flach 1983) (diagram from Miall 1985)

methods have become widely used. Some researchers have adopted the architectural classification offered in the 1996 book; most have developed variants of the classification to suit the observed characteristics of their particular field project;

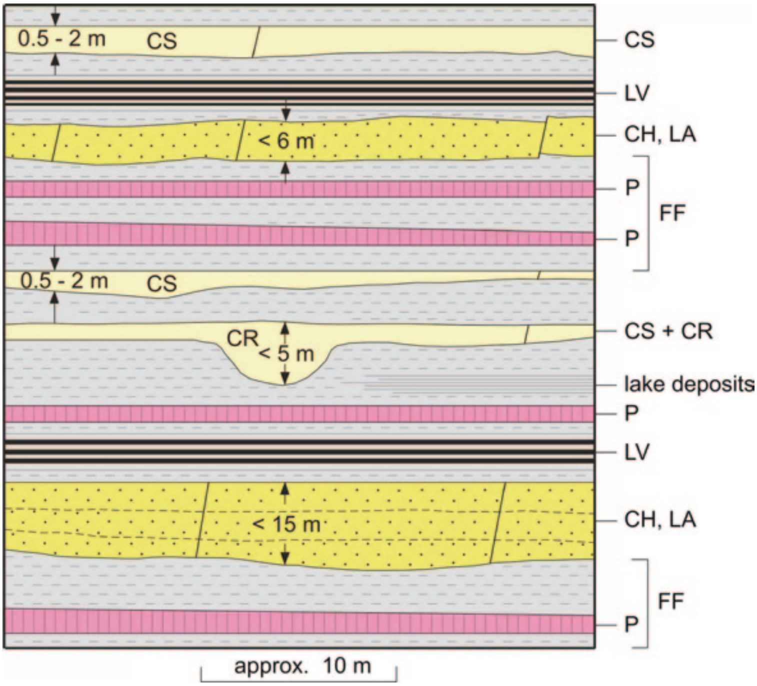


Fig. 2.36 Architectural diagram of a floodplain succession, based on the Lower Freshwater Molasse of Switzerland (Platt and Keller 1992), showing the range of elements to be expected in a floodplain setting (Miall 1996, Fig. 7.3). Element codes: *CH* channel, *CR* crevasse channel, *CS* crevasse splay, *FF* floodplain fines, *LA* lateral accretion element, *LV* levee. *P* pedogenic (paleo soil) unit

many make use of the bounding-surface classification that was also documented in the 1996 book. What follows here are some highlights of recent studies in this area.

Firstly, shown here are some general architectural classifications used for various sedimentological, stratigraphic or other purposes. Figure 2.37 illustrates an example to which the techniques of architectural-element analysis are being put, in this case the aquifer characterization of a Triassic unit in Germany. Some of the key characteristics of these elements are shown in Fig. 2.38.

Figure 2.39 illustrates an architectural classification used as the basis for a study of subsidence rates and tectonic mechanisms. López-Gómez et al. (2010) attempted to relate differences in facies architecture to differences in subsidence rate and crustal stretching factors, in a Permo-Triassic extensional basin in Spain. Sections showing the most varied architectural geometries, including ribbon and nested forms, corresponded to the highest stretching factors, reflecting tectonic phases of greater stretching and subsidence. Tectonic phases with a wider variety of fluvial geometries showed a greater difference in stretching factors, indicating stages of basin development related to different crust and lithospheric mantle activity. The field and laboratory data

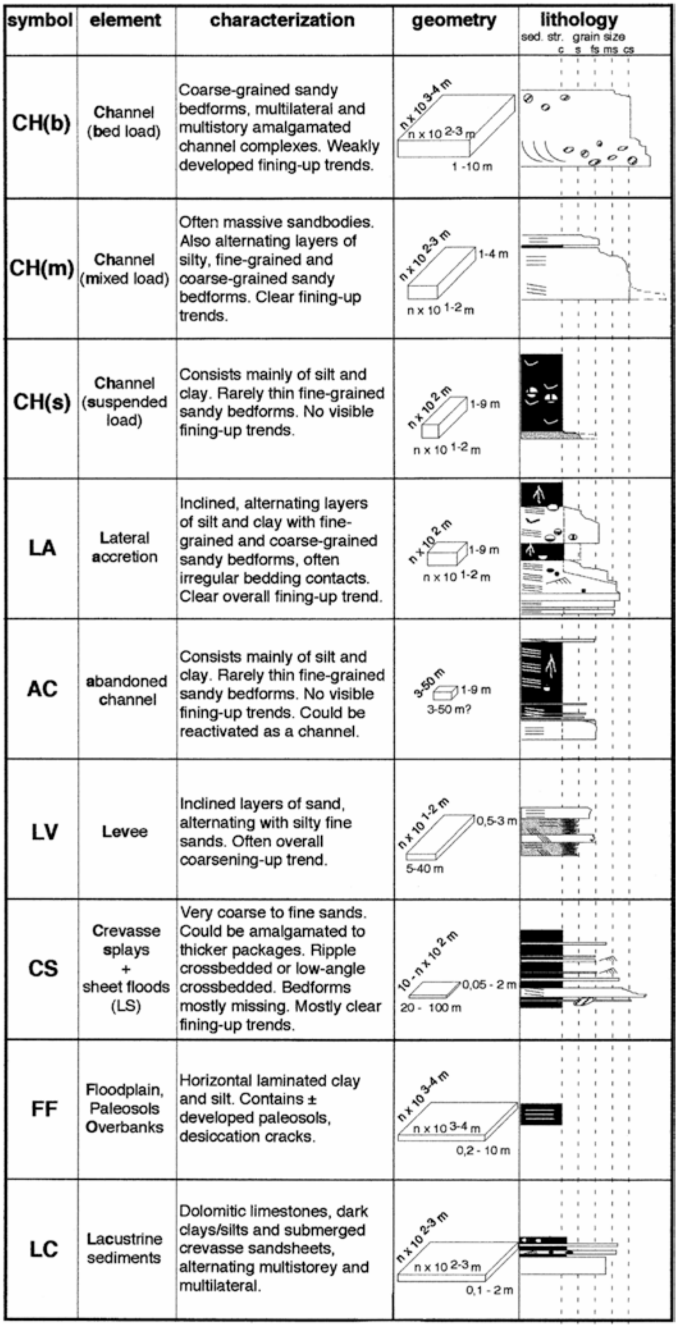


Fig. 2.37 An example of the application of architectural-element analysis to a specific field study, in this case a Triassic aquifer in Germany (from Hornung and Aigner (1999))

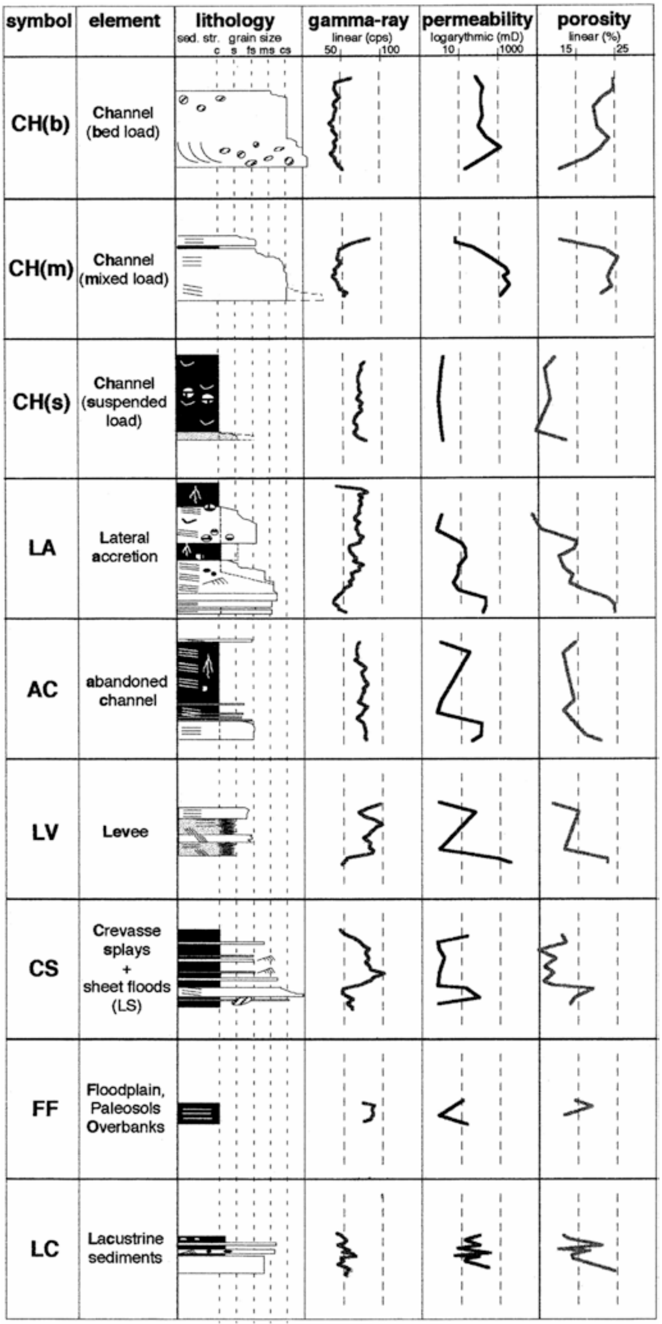


Fig. 2.38 Lithologic, wireline and porosity–permeability characteristics of the architectural elements in the Triassic aquifer studied by Hornung and Aigner (1999)

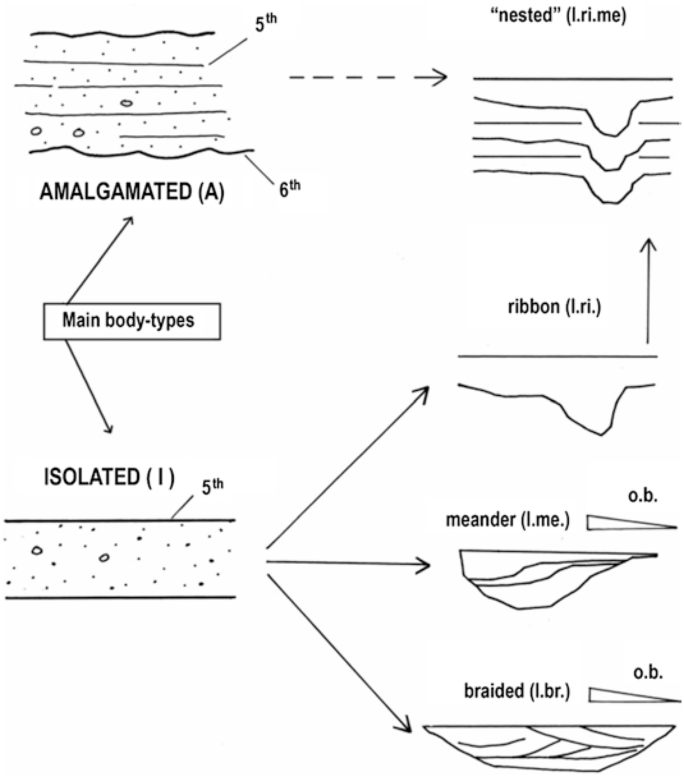


Fig. 2.39 Architectural classification of sandstone bodies in a Permo-Triassic basin in Spain (López-Gómez et al. 2010). Bounding surfaces (from Miall, 1996) are indicated by "5th" and "6th"

suggest that although general subsidence in some way controls the resultant fluvial geometry of the Permian and Triassic alluvial sediments of the Iberian Ranges, there is no simple direct relationship between the two factors. The only correlation found was between crustal and lithospheric mantle activity—reflected by their stretching factors—and fluvial geometry. It would appear that, besides subsidence, we need to consider a combination of other factors such as the rate of avulsion, climate, or budget of sediments to predict the alluvial architecture of a basin (summarized from López-Gómez et al. 2010). This study is examined further in Chap. 6.

Yuanquang et al. (2005, Fig. 10) developed a channel classification for the purpose of outcrop reservoir studies (Fig. 2.40). They carried out a detailed field program of porosity and permeability measurement and examined how the poro-perm architecture related to the facies and internal bounding surfaces of the channel systems and their component elements.

Allen and Fielding (2007) illustrated the range of architectural elements in a Permian unit in Australia (Fig. 2.41).

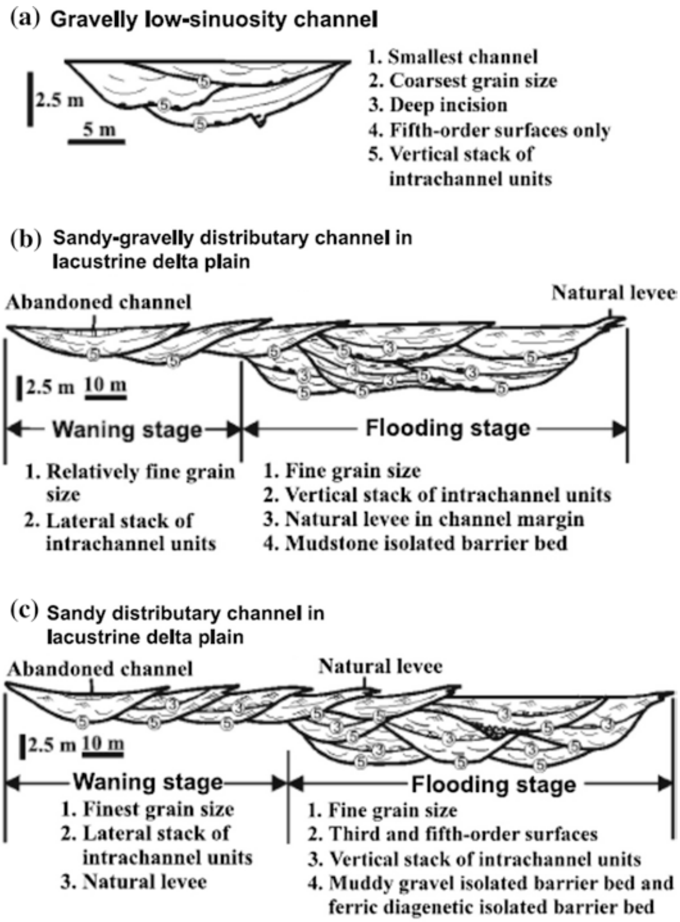


Fig. 2.40 A classification of channel systems in outcrops of a Triassic oil reservoir in western China (Yanquang et al. 2005, Fig. 10). AAPG © 2005. Reprinted by permission of the AAPG whose permission is required for further use

In a study of an ancient ephemeral system, North and Taylor (1996) developed their own classification scheme for the coarse units (many are not channelized) (Fig. 2.42), and illustrated, in a stratigraphic cartoon, how these elements form part of a fluvial-eolian succession (Fig. 2.43). These researchers chose to erect their own terminology, which may have advantages in enabling the researcher to describe unique features without assumed or implied similarities to earlier published classifications.

Long (2006) reported on a detailed study of a Paleoproterozic sandstone unit. Deposited during the pre-vegetation era, this unit is dominated by laminated sandstone sheets and sheets of sand bedforms (elements LS and SB in the Miall 1996 classification). Figure 2.44 illustrates one of the profiles from this study.

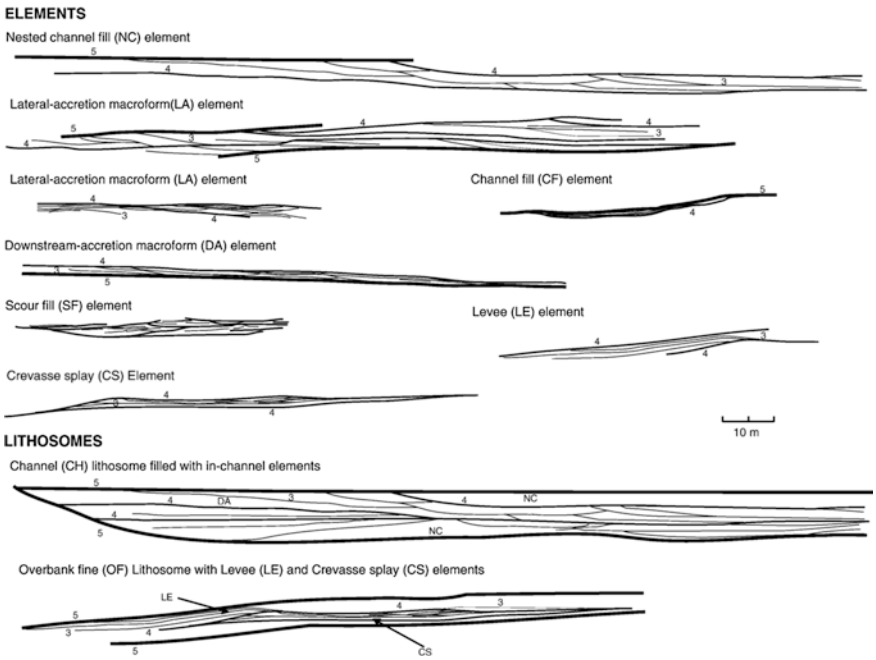
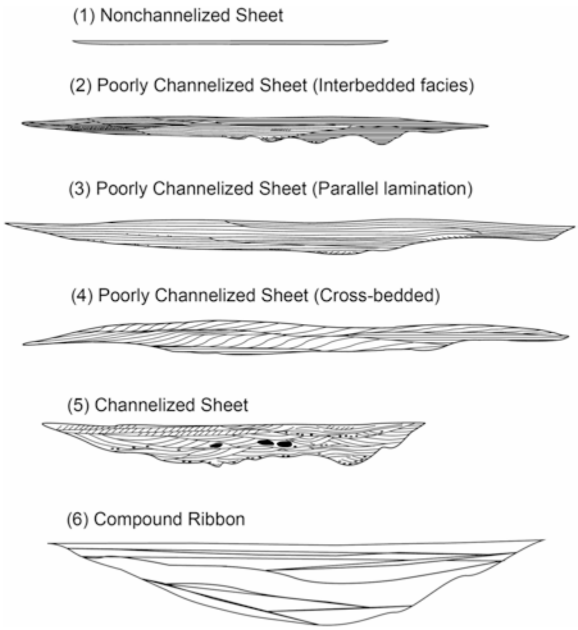


Fig. 2.41 The range of architectural elements in a Permian fluvial system in Australia (Allen and Fielding 2007, Fig. 4). Numerals refer to the bounding-surface classification of Miall (1996, 2010a)

Fig. 2.42 The elements ("facies associations") of the Kayenta Formation (Jurassic), an ephemeral system in Utah and Arizona (North and Taylor 1996, Fig. 4). AAPG © 1996. Reprinted by permission of the AAPG whose permission is required for further use



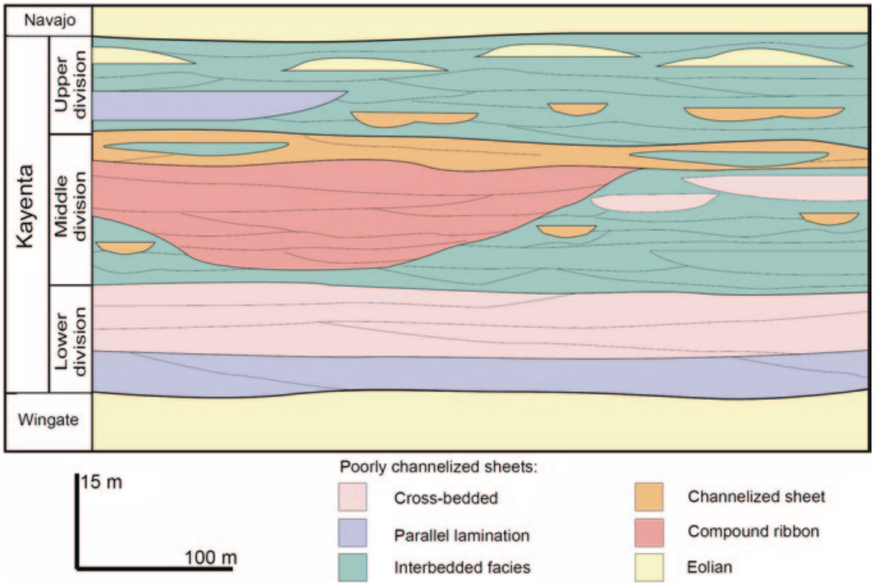


Fig. 2.43 The stratigraphic relationship of the component elements in the Kayenta Formation and the relationship of this formation to overlying and underlying eolian units (North and Taylor 1996, Fig. 6). AAPG © 1996. Reprinted by permission of the AAPG whose permission is required for further use

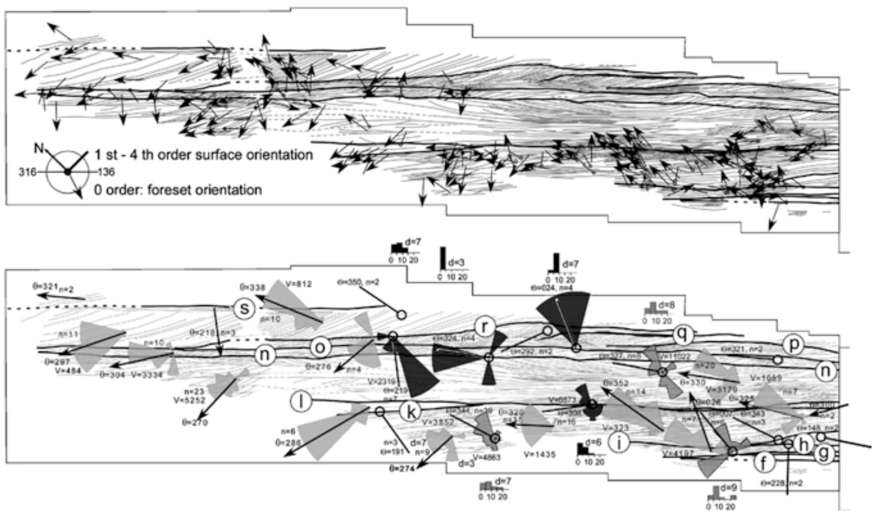


Fig. 2.44 Profile diagrams from the Paleoproterozoic Athabasca Formation Saskatchewan (Long 2006, part of Fig. 10)

Many other examples of architectural documentation and classification could be cited. The point is that the message about the three-dimensional nature of fluvial architecture has clearly been heard in the sedimentological and petroleum-geology community. There are four obvious applications of this approach:

(1) The increased ability to reconstruct convincing depositional environments provided by architecturally documented two- and three-dimensional outcrop panels.
(2) The method provides a basis for the understanding of the relationship of porosity–permeability structures in aquifers and reservoir units to primary depositional fabrics. On a larger scale, the same methods help to evaluate the nature of jigsaw-puzzle and labyrinth-type reservoir units. Application of the methods is necessary for any advanced "analog" studies of reservoirs, offering the advantage of the preserved record as a basis for comparison, rather than the commonly ephemeral features observable in modern rivers.
(3) The method is a necessary basis for studies of sequence stratigraphy in nonmarine rocks because any exploration of the dependence of fluvial architecture on allogenic controls, such as tectonism and base-level change, requires a systematic analysis of depositional architecture. Such architectural characteristics as channel-stacking patterns (for example) are clearly related to rates of change of base level or source-area uplift.
(4) Basin evolution is commonly documented using such devices as subsidence plots. This type of basic regional information may be usefully supplemented by architectural studies, which help to reveal how depositional systems respond to different patterns of basin uplift, subsidence or tilting (while bearing in mind the necessary caveats regarding time scale of accommodation generation versus the times scales of fluvial processes; see Chap. 6).

Chapter 3
Autogenic Processes: Avulsion and Architecture

3.1 Floodplain Processes

Fluvial architecture is constructed from channels and floodplains. In Chap. 2 we focused on the formation and classification of channels. In this chapter we discuss the formation of floodplains and the relationships between channels and floodplains. According to Nanson and Croke (1992), floodplains are constructed by six processes: lateral point-bar accretion, overbank vertical accretion, braid channel accretion, oblique accretion, counterpoint accretion, and abandoned channel accretion. Not all of these processes occur in every fluvial system; for example, lateral point-bar accretion and braid channel accretion tend to characterize contrasting fluvial styles, the meandering and braided style, respectively, although as noted earlier, the two processes are not entirely mutually exclusive.

The Nanson and Croke (1992) discussion and classification, although apparently dealing with floodplains, is in fact based primarily on channel style. Three broad classes of floodplain were suggested, as follows (with my own comments added). The letter classification is from their paper:

A. High-energy non-cohesive floodplains: formed mostly within confined valleys in upland regions, characterized by sudden, commonly catastrophic, cut and fill. Sheet-like deposits of coarse sand and gravel are formed primarily during flood events. Individual floodplain deposits have limited lateral extent and low preservation potential.
B. Medium-energy non-cohesive floodplains: the traditional braided, wandering and meandering fluvial systems, characterized by well-developed channel-floodplain or broad, braided channel systems.
C. Low-energy cohesive floodplains: single-thread and anastomosing systems, typified by ribbon channel bodies and broad floodplains; facies assemblages reflect arid or humid climatic conditions.

Of the six "floodplain" processes listed by Nanson and Croke (1992), only "overbank vertical accretion" would be considered a typical floodplain process, as distinct from the processes that form channel systems, and which commonly culminate in the geomorphic assemblage termed the alluvial ridge. In fact, overbank vertical accretion may be broken down into several processes that are of interest to

A. Miall, *Fluvial Depositional Systems*, Springer Geology,
DOI: 10.1007/978-3-319-00666-6_3,

sedimentologists, because of the distinct facies assemblages and fluvial architecture that result. These were listed by Brierley (1996) and constitute the following four “landscape elements”, to use his terminology (see Fig. 2.33);

Levee
Crevasse-splay
Floodplain
Backswamp.

An architectural diagram that illustrates these elements is shown in Fig. 2.36. Figure 3.1 illustrates the floodplain elements of a typical large river, a portion of the Mississippi system.

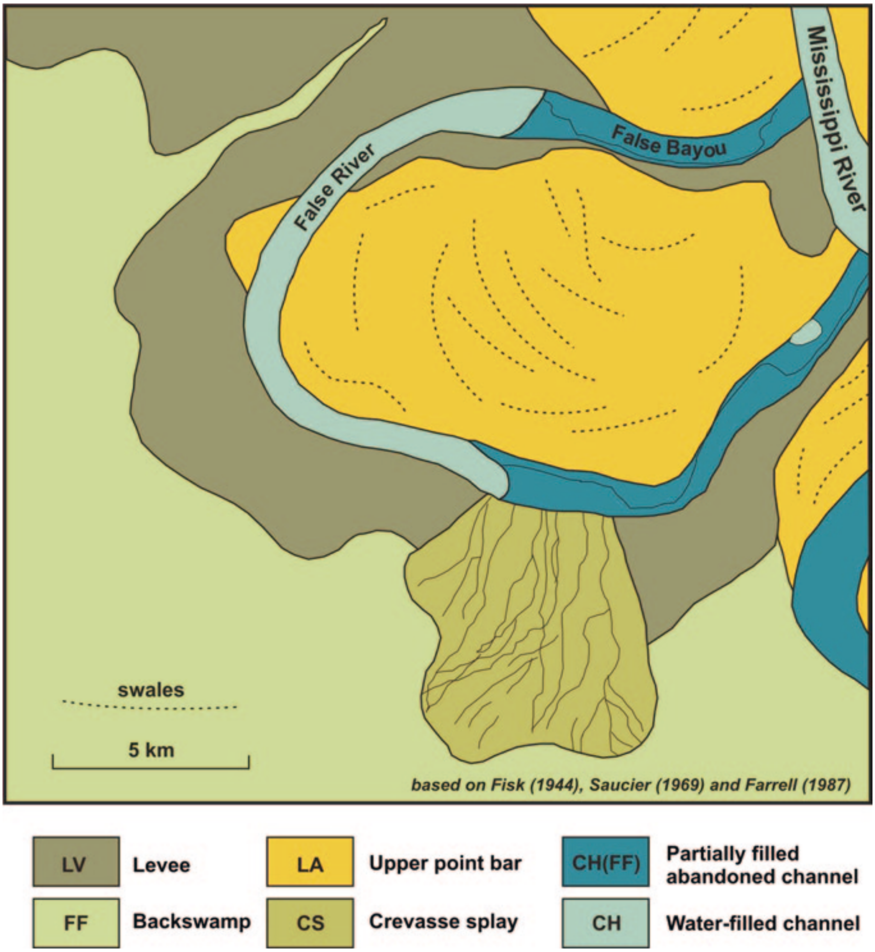


Fig. 3.1 The landscape elements and deposits of a floodplain, showing part of the Mississippi system (after Farrell 1987)

Floodplain construction involves repetition of various autogenic and allogenic processes over a range of time scales. Individual flood events, bar deposition and migration, and lateral accretion, are primarily autogenic processes and construct discrete depositional units over time scales of up to a few years. Kraus and Wells (1999) referred to these as micro-scale (<1 m thick, duration of days to months) to meso-scale (>1 m thick, 1–10^2 yr duration) events. They form metre-scale cycles, termed *fluvial aggradation cycles* (FAC) by Atchley et al. (2004). Longer-duration cycles, termed macro-scale (>10 m thick, 10^3–10^4 yr duration), and mega-scale (>100 m, 10^5-10^7- yr duration) cycles, develop stacked cycles, termed FAC-sets (Atchley et al. 2004), and are the product of allogenic processes.

The most important floodplain process is *avulsion*, the shifting of channels into new positions on the floodplain. This is usually described and discussed in terms of a sudden process. The word has origins in medical terminology. A Wikipedia definition of the term is that avulsion is "a form of amputation, where the extremity is pulled off rather than cut off." The online Free Dictionary refers to avulsion as "The forcible tearing away of a body part by trauma or surgery." As Mohrig et al. (2000, p. 1787) stated: "River avulsion is the relatively rapid transfer of river flow out of an established section of channel belt and into a new flow pathway elsewhere on the flood plain. In river systems where avulsions occur, the process typically dominates the long-term dispersal of sediment and water across their alluvial surfaces (Fig. 3.2). As a result, understanding and characterizing the controls on avulsion are important to the geomorphologist studying evolution of depositional landscapes, to the sedimentologist reconstructing the history of ancient river systems, and to the civil engineer interested in controlling the position of waterways."

Avulsion is typically initiated by localized erosion of a channel bank, forming a *crevasse channel*, which diverts some of the discharge and sediment load from the main channel onto the floodplain. Initiation of crevasse erosion depends on a number of factors, to be discussed below. Continued erosion within the crevasse channel will deepen and widen it, and may lead to the formation of a *crevasse splay*, a type of mini-delta splaying out from the channel bank into the floodplain. Figure 3.1 illustrates a large, compound crevasse splay formed from the deposits of a number of small crevasse channels that developed along the banks of the

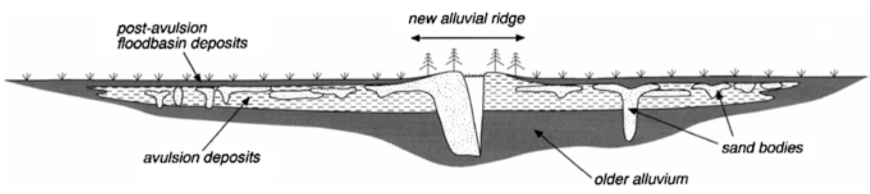


Fig. 3.2 Idealized model of an alluvial ridge, with an active channel flanked by levees and by avulsion deposits. The latter are mainly composed of silt and clay, with sand bodies deposited in the minor channels of crevasse splays. From Makaske (2001, Fig. 10, p. 173), based on the work of Smith et al. (1989) in the Cumberland Marshes of Saskatchewan

Mississippi River, and Fig. 3.2 illustrates the typical architecture of the crevasse splay deposits within avulsion complexes. Under certain circumstances, a crevasse may eventually cause the diversion of most or all of the discharge passing down the main channel, with the result that the main flow of the river is diverted into a new course. An analysis of why this happens, and what are the consequences for fluvial deposition, has consumed much energy on the part of fluvial sedimentologists. Why is it regarded as an important question? Because it appears to be the key to the understanding of how large-scale fluvial architectures are constructed. Viewed in their simplest form, fluvial successions consist of stacked channel deposits separated from each other by floodplain fines. The nature of the stacking and of the interconnectedness of the channel bodies, is the key to understanding fluid flow in the subsurface, and is therefore of critical interest to hydrogeologists and reservoir geologists.

An assumption that has exerted a major influence on interpretations and models is that given a constant avulsion frequency, the density and connectedness of channel sand bodies should be inversely proportional to the aggradation rate (Bridge and Leeder 1979). Is avulsion frequency typically constant? What controls avulsion events?

As Mohrig et al. (2000, p. 1787) noted:

> Recent numerical models that simulate construction of the depositional architecture (channel stacking pattern) of fluvial deposits have emphasized the role of avulsions as a control on the distribution of ancient river channels in space and time (Allen 1978; Leeder 1978; Bridge and Leeder 1979; Mackey and Bridge 1995; Heller and Paola 1996). Key assumptions of these models concern (1) when a channel will avulse, and (2) the position of the new channel belt following avulsion. In the first assumption, it is essential to know if there is some single aspect of the channel system that primarily, and predictably, creates conditions favorable for avulsion. If so, this parameter may serve as a standard among systems so that different-sized rivers can be compared. The second assumption evaluates whether preexisting topography exerts a role on site selection of the newly formed stream. Some workers have suggested that former channel positions create alluvial ridges along the flood plain that repel new channels (Allen 1978; Bridge and Leeder 1979; Mackey and Bridge 1995). In contrast, studies of modern avulsions show that newly avulsed rivers are strongly attracted toward preexisting channels, the beds of which are preserved as lows on the flood plain (Aslan and Blum 2000; Morozova and Smith 2000). Stratigraphic evidence for this attraction was presented by Maizels (1990).

Mohrig et al. (2000, p. 1788) observed that the process of river avulsion can be considered as having two fundamental requirements: “a setup, where the river aggrades over tens to thousands of years and becomes poised for avulsion, and a trigger, which is a short-term event that causes the slow or abrupt abandonment of the channel” (typically floods; Mosley 1975; Brizga and Finlayson 1990; Slingerland and Smith 1998). River avulsion setup is generally recognized to be a consequence of the tendency for deposition rates near river channels to be greater than those on the adjacent flood plain. The coupled deposition of sediment on the bed of a channel and on the levees flanking the channel margins leads to the channel becoming progressively perched above its flood plain while maintaining a cross-sectional shape that is best suited for the throughput of water and sediment (Mohrig et al. 2000, p. 1788).

According to Jones and Schumm (1999) there are four major groups of processes that trigger avulsion (Table 3.1). The first two of these groups focus on changes in the relative slopes of the main channel and the potential avulsion channel as a result of sedimentation. As they noted (Jones and Schumm 1999, p. 172), "avulsions are not always triggered by the largest floods on a given river" because it may take an extended period for sedimentation to build the slope differentials to a critical threshold value. The construction of an alluvial ridge flanking and containing the main active channel is in the long-term (geomorphically speaking) an unstable condition that eventually is likely to lead to avulsion. Tectonic activity, such as tilting of the valley floor, or movement on a fault that crosses the floodplain, may also trigger avulsion. The most well-known example of the first process is the study of the South Fork Madison River in Montana by Leeder and Alexander (1987).

Table 3.1 Causes of avulsion (Jones and Schumm 1999)

Processes and events that create instability and lead toward an avulsion threshold and/or act as avulsion triggers		Can act as trigger?	Ability of channel to carry sediment and discharge
Group 1. Avulsion from increase in ratio, S_a/S_e owing to decrease in S_e	a. Sinuosity increase (meandering)	No	Decrease
	b. Delta growth (lengthening of channel)	No	Decrease
	c. Base-level fall (decreased slope[a])	No	Decrease
	d. Tectonic uplift (result in decreased slope)	Yes	Decrease
Group 2. Avulsion from increase in ratio, S_a/S_e owing to increase in S_a	a. Natural levee/alluvial ridge growth	No	No change
	b. Alluvial fan and delta growth (convexity)	No	No change
	c. Tectonism (resulting in lateral tilting)	Yes	No change
Group 3. Avulsion with no change in ratio, S_a/S_e	a. Hydrological change in flood peak discharge	Yes	
	b. Sediment influx from tributaries, increased sediment load, mass failure, eolian processes	Yes	Decrease
	c. Vegetative blockage	No	Decrease
	d. Logjams	Yes	Decrease
	e. Ice jams	Yes	Decrease
Group 4. Other avulsions	a. Animal trails	No	No change
	b. Capture (diversion into adjacent drainage)	–	No change

Notes: S_a is the slope of the potential avulsion course, S_e is the slope of the existing channel
[a]In settings where the up-river gradient is greater than the gradient of the lake floor or shelf slope, base-level fall may result in river flow across an area of lower gradient

As discussed below, avulsion typically results in the formation of channel clusters, as viewed in stratigraphic strike-oriented cross-section. An understanding of the processes which lead to channel clustering and the stacking or displacement of clusters through stratigraphic time is essential for the understanding of channel-body architecture and connectivity.

Undoubtedly the most detailed study yet carried out on avulsions in a modern river setting is that of the Rhine Meuse system. A very large data base of drill holes and ^{14}C dates has permitted a very detailed reconstruction of the history of this river and delta system since the last glaciation ended, and fluvial processes commenced about 8000 BP. This is referred to below. Stouthamer et al. (2011, p. 225) stated that:

> Local/regional aggradation rate ratio and CDP [channel deposit proportion] are positively related. The balance between local and regional aggradation rates determines cross-valley gradients, which is generally regarded [as] an important prerequisite for avulsion. In this view, the ratio of local to regional aggradation rate is a proxy for the probability of avulsion.

They confirmed the importance of what is termed the "*superelevation* of the channel or the channel belt above the floodplain and the associated increase in cross-valley slope relative to either the downchannel or the downvalley slope." (op. cit., p. 226).

There have historically been four ways to approach the problem of understanding avulsion:

(1) Historical documentation of avulsions in existing modern river systems (maps, ^{14}C dates of avulsion deposits, etc.);
(2) Numerical models of avulsion based on geometrical simulations of floodplain processes;
(3) Scale modeling of fluvial processes in laboratory tanks;
(4) Detailed studies of the facies and architecture of pre-Recent and ancient avulsion successions directed towards understanding the specific local processes at work.

3.2 The Historical Record of Avulsion in Modern Rivers

Fluvial geomorphologists have for some time studied the process of avulsion in modern rivers. Very few case studies of actual avulsion events are available (Bryant et al. 1995, p. 365), but historical records, including the reconstruction of avulsion histories from old maps, have provided much information. For such major fluvial systems as the modern Mississippi River (Saucier 1974, 1994), Yellow River (Hillel 1991; Li and Finlayson 1993), Po River (Nelson 1970), Rhine-Meuse system (Törnqvist 1994; Stouthamer 2001a, b; Berendsen and Stouthamer 2001; Hesselink et al. 2003; Stouthamer et al. 2011), and the Kosi river in India (Wells and Dorr 1987; Singh et al. 1993), the history of avulsion over the last few hundreds

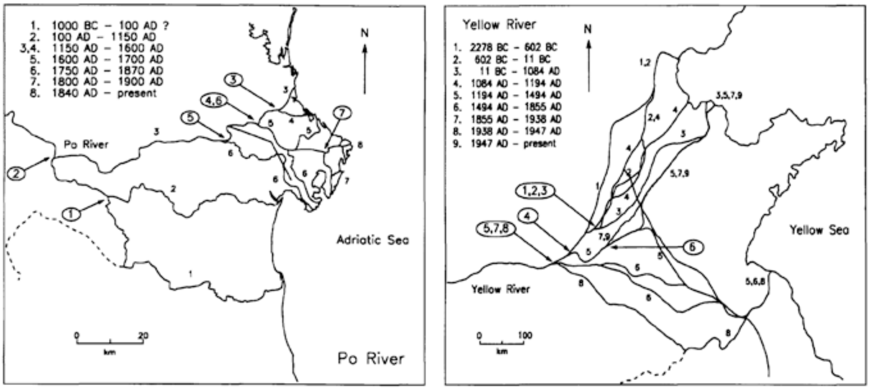


Fig. 3.3 Avulsion history of the Po River, Italy, and the Yellow River, China (Mackey and Bridge 1995)

or thousands of years has been well documented (Figs. 3.3, 3.4). These rivers have, at various times and by various authors, been used as examples on which to base geological models of the avulsion process, for use in unraveling the origins of avulsion architectures in the ancient record (e.g., Mackey and Bridge 1995). The usefulness of these documented histories may be of limited value, because they are based on recent processes with a very brief record of stratigraphic evolution, and one that is characterized by the uniqueness of a post-glacial history of a recent rapid sea-level rise, with all the consequences of that for a rapid increase in accommodation in coastal regions.

The Mississippi and Rhine-Meuse systems have been explored extensively using numerous drill holes, and the stratigraphic record of fluvial evolution is particularly well known for these major systems. However, the caveat regarding the possible uniqueness of the post-glacial period in earth history remains. Törnqvist (1994) measured variations in overall avulsion frequency in the Rhine-Meuse Delta in the Netherlands by ^{14}C methods. He found a higher rate of avulsion in the delta system during the interval 8500 BP–4300 BP than from 4300 BP to the present. The high-frequency interval corresponds to a period of faster sea-level rise and was thus inferred by Törnqvist to represent a period of increased fluvial aggradation rate. Stouthamer et al. (2011) updated these observations on the basis of much additional data (Fig. 3.5). They documented a high avulsion frequency from 8500 to 7300 BP, with a drop in frequency following, until 3200 BP., possibly related to "the steadily dropping rate of accommodation growth under decreasing rates of sea-level rise" (Stouthamer et al. 2011, p. 222). An increase in avulsion frequency after 3200 BP is associated with an increase in the fine sediment load in the river system, and occurred despite the much slower rate of accommodation generation associated with the slower rate of sea-level rise during this period. We return to these river systems below.

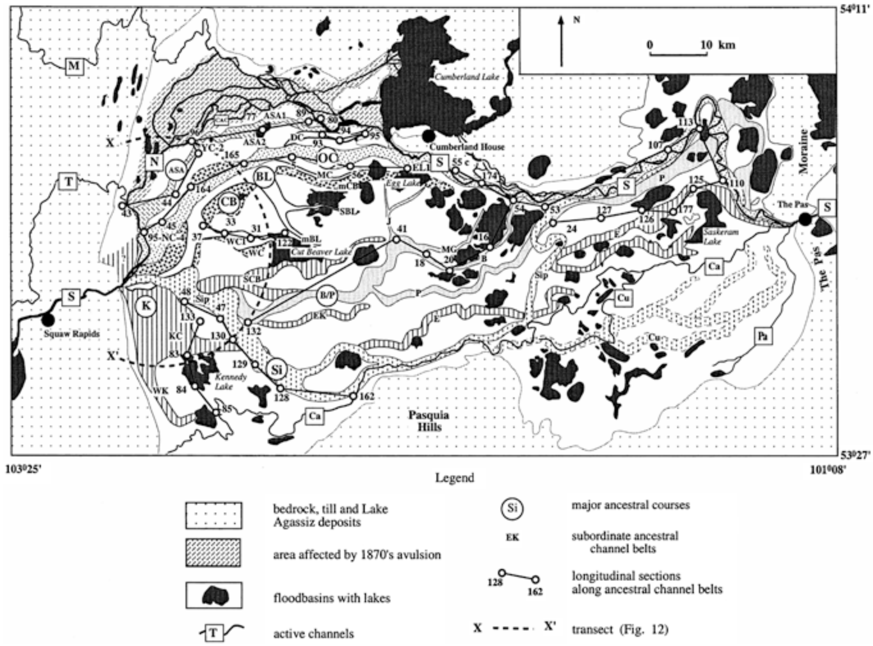


Fig. 3.4 Major geomorphic elements of the Cumberland Marshes (Morozova and Smith 1999, Fig. 2, p. 233)

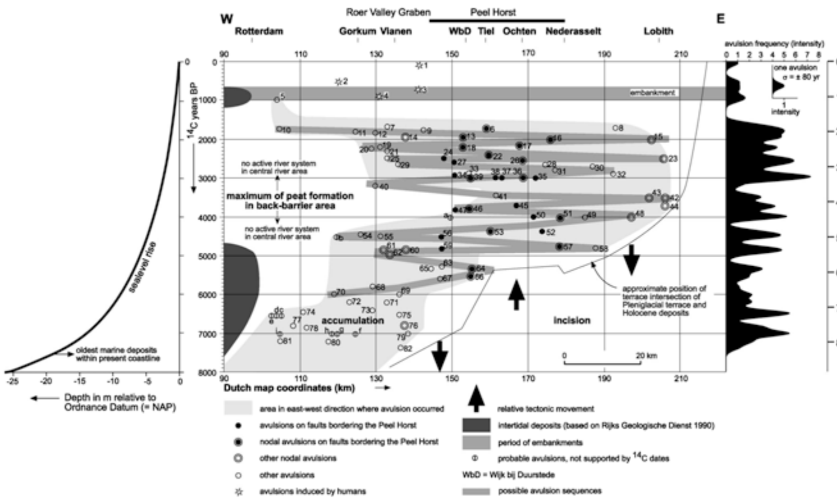


Fig. 3.5 Avulsion history of the Rhine River (Stouthamer et al. 2011, Fig. 6, p. 223)

In a study of the modern Paraná River in Brazil, Stevaux and Souza (2004, pp. 63–64) stated: "Although the average anastomosing process is related to rapid sediment aggradation, vegetation (Smith and Smith 1980), bank cohesiveness and low stream power, and this system can occur in a range of climatic environments (Nanson and Croke 1992), in the present case climatic change and subtle tectonic influence are suggested." They noted that during periods of climate change the river system underwent a period of instability corresponding to a short erosion interval. They cited a short period of semi-aridity in central-southern Brazil and eastern Argentina in the Upper Holocene (ca. 3.6–1.5 ka BP) during which time a period of alluvial fan construction could be recognized, the channel thalweg underwent incision, and eolian erosion and dune sedimentation took place. They suggested that the identified avulsion events originated from changes in river hydrology.

The Baghmati River, India, locally displays a pattern of avulsion that can throw some light on the processes of architectural development. This river has been described as "hyperavulsive" by Sinha et al. (2005, p. 192). Eight major and several minor avulsions have been documented over the last 250 years, indicating a decadal avulsion frequency. Several different types of avulsion occur, including channel cutoffs, abandonment by floods, and nodal avulsions. Most avulsions are local; many channels may be active simultaneously. The initiation of new channels or old channel reoccupation during flood events appears to be the dominant process, the regular occurrence of high-discharge events during the annual monsoon likely being a major factor in the evolution of the system.

Jones and Hajek (2007, pp. 133–134) observed that research on modern systems has noted a difference in association of river systems and river avulsion with crevasse splay deposition:

> Aslan and Autin (1999) suggest that floodplain aggradation is dominated by crevasse splay deposition during intervals of active avulsion of the Holocene lower Mississippi River. Kraus and Wells (1999) attribute up to 50 % of floodplain aggradation in the Willwood Formation to avulsion-related non-overbank deposition. This is similar to the percentage of crevasse splay deposition during the Holocene lower Mississippi River avulsions. Such settings are likely to produce stratigraphically transitional avulsion stratigraphy. In contrast, there are other modern river systems known to be much less splay-prone. Studies in several modern river systems, such as Narew River (NE Poland), Okavango fan rivers (Botswana), Fitzroy River (Australia) and Cooper Creek (Australia) have documented river avulsions that do not fit the Smith et al. (1989) avulsion model of river avulsion (Rust and Legun 1983; McCarthy et al. 1992; Taylor 1999; Knighton and Nanson 2000; Aslan et al. 2003; Gradzinski et al. 2003; Fagan and Nanson 2004).

The rivers listed above have avulsed by developing new channels by nick-point retreat and/or overland flow on the floodplain. The modern Fitzroy River, northwest Australia (Taylor 1999) demonstrates that channel avulsion is not associated with crevasse splay deposition in that system (Jones and Hajek 2007). Rather, channels relocate into topographic depressions on the floodplain, which are generated by scouring during overbank flow. Taylor (1999) suggested that crevasse splay facies should be rare or absent in the Fitzroy's stratigraphic record, in part because of levees with very low topographic relief (800 m wide by <1 m high), high-magnitude flooding events, and highly cohesive floodplain material. Such a system could dominantly produce stratigraphically abrupt channels. In comparison, Ferris Formation channels lack well-developed levee deposits and have muddy floodplains, which might indicate low-relief levees and highly cohesive banks, respectively (Jones and Hajek 2007, p. 134).

Sambrook-Smith et al. (2010) demonstrated that scour depths and channel and bar features developed during a major, 40-year flood in the braided Saskatchewan River, Canada differed in no major way from those features of the fluvial morphology that were modified during normal, annual bankfull floods, making it difficult to distinguish the effects of discharge variability in rivers of this type. They noted similar conclusions having been reached in several other large river systems. These observations also have relevance for the concept of geomorphic threshold, as applied to the triggering of major changes (including avulsions) in river systems.

3.3 Avulsion in the Recent Geological Record

The first detailed studies of the recent stratigraphic record of avulsion were carried out in the Mississippi system (Saucier 1974, 1994; Autin et al. 1991; Asland and Autin 1999; Aslan et al. 2005), the Cumberland Marshes of Saskatchewan (Smith et al. 1989), and the Rhine-Meuse system (Törnqvist 1993, 1994). These studies have shown that overbank sheet flooding is not a major mechanism for the accumulation of floodplain deposits. Most floodplains are built by the development of crevasse channels, which divert water and sediment into crevasse splays that, in turn disperse sediment and water primarily into low standing areas of the floodbasin, in particular into preexisting, abandoned channels. The reoccupation of abandoned channels has emerged as a very significant process in floodplain development, which means that channel sandstone bodies are, in many cases, multistory units representing several phases of sedimentation.

In the Cumberland Marshes of Saskatchewan, a major avulsion was triggered by an ice jam in 1873. The Saskatchewan River system branches into numerous channels in this area, and the event studied by Smith et al. (1989) represents only the latest of a series of avulsions that have constructed the floodplain stratigraphy since fluvial sedimentation began to fill the post-glacial Lake Agassiz some 8,500 years ago. A detailed investigation of the shallow subsurface stratigraphy provided the basis for a general model of avulsion and stratigraphic evolution (Smith et al. 1989). An initial crevasse feeds a splay deposit that gradually grows larger and more complex, with a system of interconnected, anastomosing channels occupying an ever larger area of the floodbasin. Eventually, a single channel becomes dominant because of a small local slope advantage. It begins to incise earlier splay deposits, and also begins to develop meanders. Channel aggradation and point-bar development along this dominant channel then combine to build a new alluvial ridge, setting the stage for build-up to the next avulsive event when threshold conditions are reached again.

In a later study, Morozova and Smith (1999, 2000) examined earlier episodes of avulsion stored in the stratigraphic record of the Cumberland Marshes. They

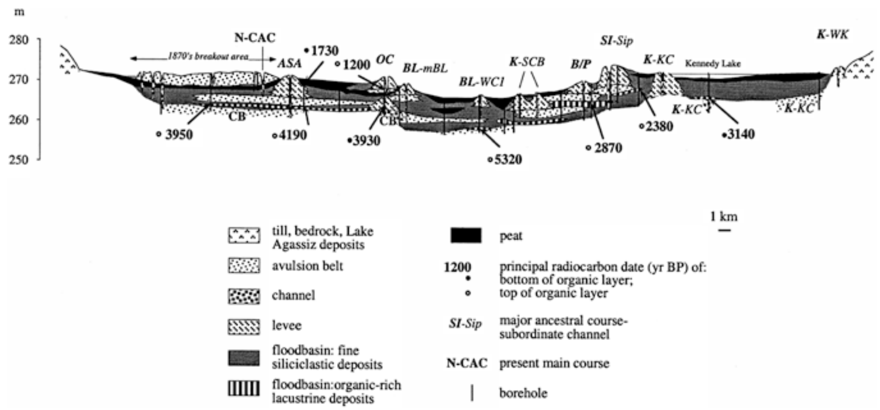


Fig. 3.6 Cross-section through the Cumberland Marshes, along the transect X–X′ of Fig. 3.4 (Morozova and Smith 2000, Fig. 2, p. 84)

documented nine major avulsions in the last 5,400 years, averaging one event every 600 years (Fig. 3.4). The life-span of the avulsion complexes overlap each other, indicating that new avulsions were accompanied by gradual abandonment of the earlier channels. Most of the avulsions were nodal avulsions, that is, they occurred near the entrance to the Cumberland Marshes, where river slope decreases abruptly. Some of the avulsions led to the reoccupation of earlier channels, others filled floodbasins by the lateral progradation of splay complexes. A cross-section through the basin exhibits a network of partially intersecting splay complexes consisting of narrow channels and levees overlying broad splay complexes that in most cases are several kilometers across (Fig. 3.6).

In the Rhine-Meuse system, avulsion event periodicity during the last 8 millennia increased from 500 years at 8,000 BP to 1,500 years at 2,800 BP and then decreased to 1,000 years at 1,500 BP (Stouthamer et al. 2011; Fig. 3.5 of this book). The frequency was greater in the immediate post-glacial period, when sea-level was rising rapidly. The river assumed an anastomosing style during this period, with a gradual change to a meandering style as the rate of sea-level rise, and the rate of avulsion both decreased after 4,300 yr BP. Törnqvist (1994, p. 714) stated:

> The number of active Rhine distributaries in the Rhine-Meuse delta increased from about four (8000 cal. yr BP) to a maximum of ten (5500 cal. yr BP) and then gradually decreased to about five until the late medieval embankments reduced the number to three. The period with the highest number of distributaries (6000–3500 cal. yr BP) coincides well with the period of dominance of anastomosing fluvial style (Tornqvist 1993a).

The Mississippi system is particularly interesting because of the incipient avulsion of the main river channel into the Atchafalaya river channel, a process that was well underway by the early 1950s, until brought under control by the U.S. Army Corps of Engineers in order to protect the flow passing down channel through New Orleans, a vital port city.

The avulsion history of the lower Mississippi is documented by Aslan et al. (2005). During the late Holocene the Mississippi River avulsed three times, with the Mississippi–Atchafalaya diversion representing a fourth incipient avulsion (Figs. 3.7–3.9). Until about 500 years ago the Mississippi River flowed on the western side of the valley, terminating in the Teche subdelta (Saucier 1994). An avulsion event near Old River, shifted the river course to the eastern side of the valley (Avulsion #1 in Fig. 3.7). By 1,500 yr BP, the Lafourche subdelta became active and prograded across the abandoned Teche subdelta. Avulsion upstream near Vicksburg, Mississippi (Avulsion #2, north of the area shown in Fig. 3.7) completed the eastward shift of the Mississippi River. Avulsion near the Lafourche avulsion site (Avulsion #3 in Fig. 3.7) led to the development of the modern Balize subdelta. By the mid-eighteenth century a switch to the Atchafalaya River was underway.

Fisk (1952) suggested a four-step evolution for the Atchafalaya River (Fig. 3.8). As described by Aslan et al. (2005, p. 653):

> Initially, the Mississippi and Red rivers flowed south along separate courses and joined downstream of Old River (Fig. 3.8a) A westward migrating meander of the Mississippi River, Turnbull Bend, captured the Red River (Fig. 3.8b). Crevassing along the western edge of the meander initiated the development of the Atchafalaya River in the Eighteenth

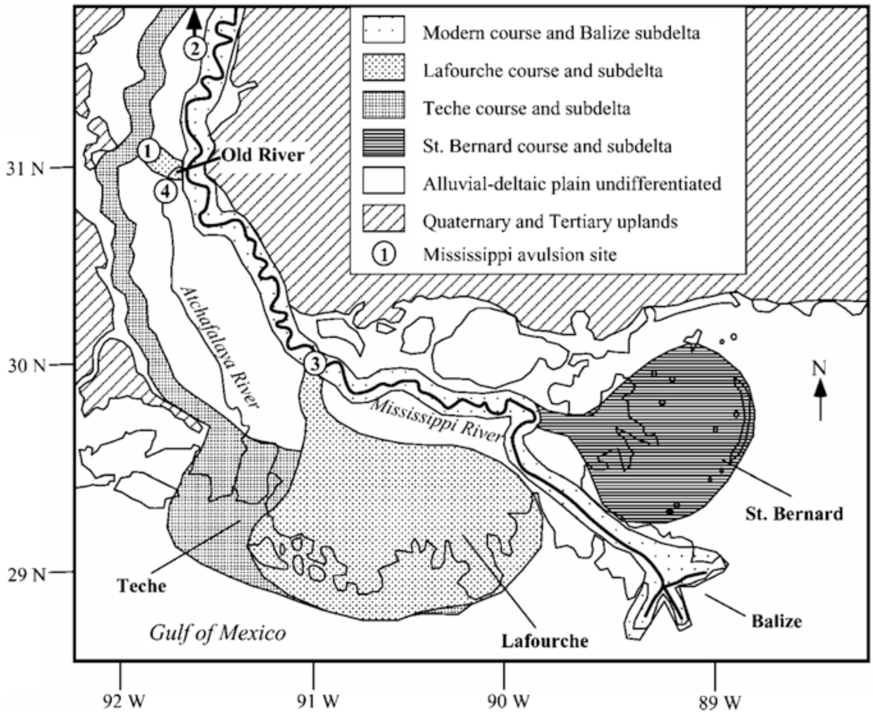


Fig. 3.7 Late Holocene Mississippi River courses, subdeltas, and locations of avulsion nodes (*numbered circles*). The avulsions are numbered from 1 to 4 (Aslan et al. 2005, Fig. 3, p. 652)

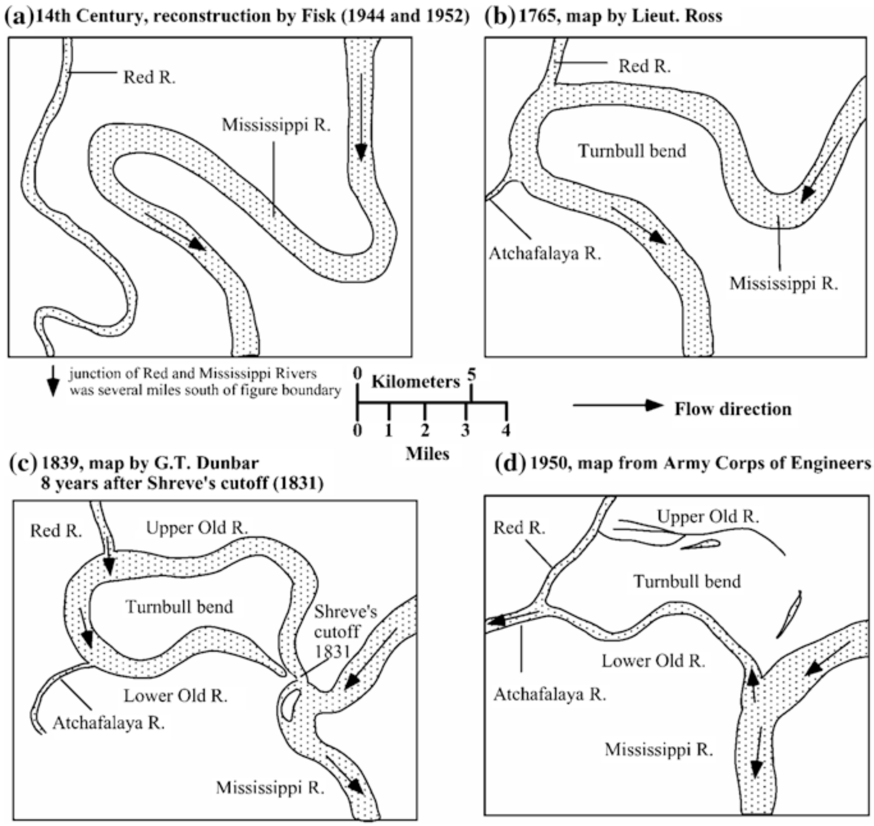


Fig. 3.8 Development of the Atchafalaya River. Modified from Fisk (1952) (Aslan et al. 2005, Fig. 4, p. 653)

Century (Fisk 1952), and by 1765, the Atchafalaya River was well established (Fig. 3.8b). In 1831, an artificial cutoff (Shreve's cutoff) across Turnbull Bend minimized flow between the Mississippi and Atchafalaya rivers (Fig. 3.8c). Subsequent dredging of Lower Old River between the 1880s and 1930s maintained flow between these two major waterways. Upper Old River filled with sediment during this time, which led to the capture of the Red River by the Atchafalaya River and its continued enlargement (Fig. 3.8d). By 1950, the Atchafalaya was transporting; 25 % of the Mississippi River discharge, and the growth of the Atchafalaya River and its down-valley gradient advantage over the Mississippi made clear that capture was imminent (Fisk 1952). Ensuing construction of the Old River Control Structures by the U.S. Army Corps of Engineers has, at least temporarily, arrested this avulsion.

Guccione et al. (1999) studied an abandoned channel in the upper Mississippi valley. An avulsion event had occurred as a result of a within-channel slope advantage, but was abandoned when the avulsion node from which it developed was cut off when the meander on which it was located became cut off and abandoned.

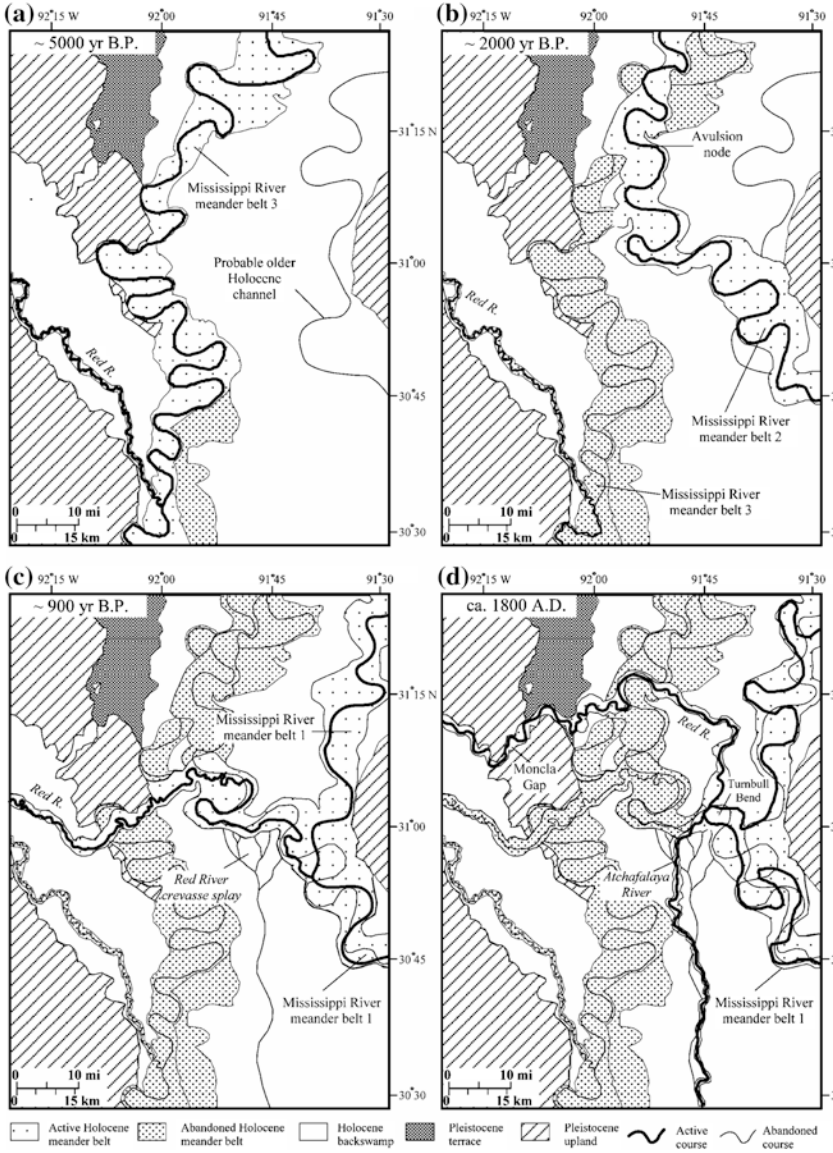


Fig. 3.9 Floodplain development and avulsion history in the Old River area over the past ~5,000 years (Aslan et al. 2005, Fig. 11, p. 660)

The Jamuna (Brahmaputra) River of Bangladesh is currently the world's largest active braided river system, and has been much studied as a possible analog for ancient sandy braided systems (Coleman 1969; Miall and Jones 2003). Since

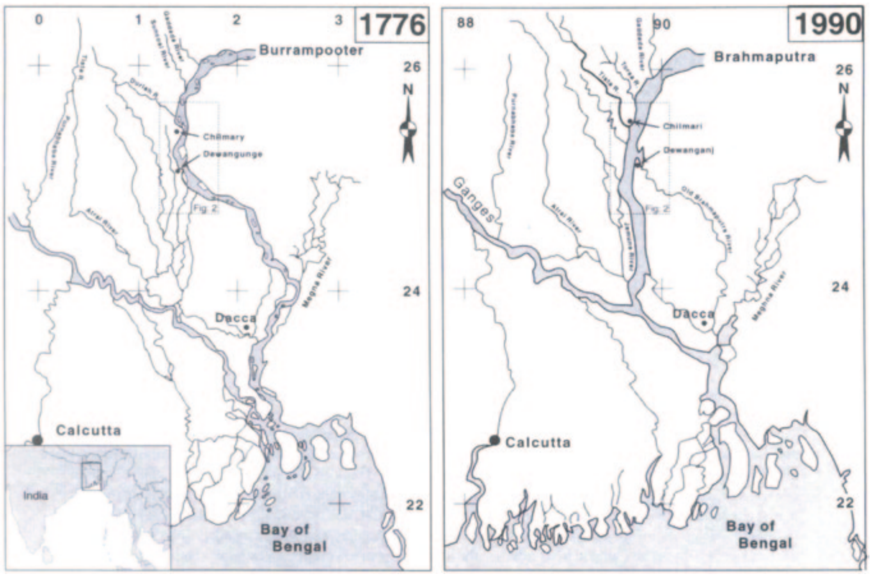


Fig. 3.10 The evolution of the Jamuna (Brahmaputra) river system, Bangladesh. Maps redrawn from original sources by Bristow (1999, Fig. 1, p. 222)

the late eighteenth century the main course of the river, north of Dacca, has shifted some 100 km westward. The town of Dewanganj, formerly on the west bank of the river, is now on the east bank (Fig. 3.10). Early work, such as that by Morgan and McKintire (1959) had suggested that tectonic tilting or river capture may have been the cause of this major shift, but Bristow (1999) makes a convincing case that existing map and other historical evidence does not support these ideas, and that the shift was entirely autogenic in origin. Figure 3.11 illustrates the evolution of the Brahmaputra and its tributaries in the area of Dewanganj, in northern Bangladesh. Several old river courses diverge from the main river near and downstream from this town, and it seems likely that the major diversion began simply as a case of "flow divergence around the mid-channel islands, with one channel directed towards the west (right) bank, caused bank erosion, which led to flow diversion into an existing floodplain channel, and which the Brahmaputra exploited and enlarged to form the Jamuna" (Bristow 1999, p. 225). This process of simple autogenic channel and bar development led eventually to this dramatic shift in the course of the river, and constitutes a good example of how autogenic processes may have substantial effects on the geography of the river system. The shifting of the Mississippi, Po and Yellow River deltas (Fig. 3.3) are other examples. The implications of these processes for the basin-filling process are clear.

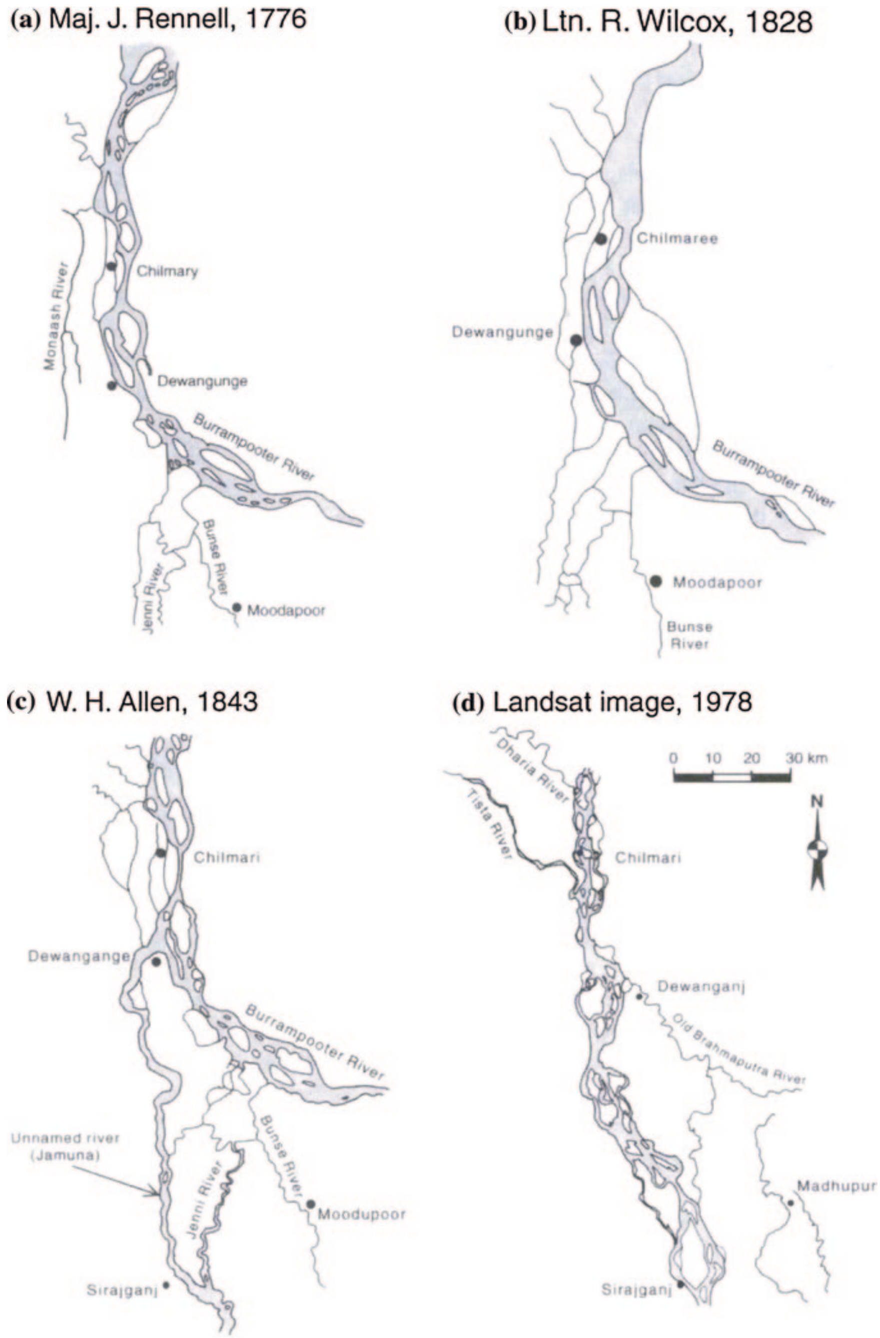


Fig. 3.11 The evolution of the Jamuna (Brahmaputra) river system, northern Bangladesh. Maps redrawn from original sources by Bristow (1999, Fig. 2, p. 224)

3.4 Avulsion in the Ancient Geological Record

Another way to study avulsion is to examine the preserved geological record. This may provide for a longer record of avulsions, through the examination of the details of a lengthy stratigraphic record, at the expense of an accurate record of timings and rates. This is an approach employed by several groups of researchers. We discuss here the results of several such studies.

Kraus and Wells (1999) reviewed criteria for recognizing avulsion deposits in the ancient record. What they termed "heterolithic avulsion deposits" commonly underlie the main channels. Heterolithic avulsion deposits are laterally extensive, pedogenically modified, fine-grained deposits, in which narrow ribbon sandstones (width:depth <10) and thin sheet sandstones occur, with paleoflow parallel or subparallel to the trunk channels. Heterolithic avulsion deposits may also be interbedded with coal or moderately well developed to well-developed paleosols. As Jones and Hajek (2007, p. 125) noted, these observations match depositional successions documented in modern avulsions, particularly the Saskatchewan River (Smith et al. 1989). Because of this convincing connection between ancient and modern, the Kraus and Wells (1999) model has become a popular way to interpret ancient alluvial stratigraphy.

Mohrig et al. (2000) examined extensive outcrops records in a portion of the Oligoccene fill of the Ebro Basin in northern Spain (Guadalope-Matarranya System), and the Eocene Wasatch Formation of western Colorado. Many of the channel fills in the Guadalope-Matarranya System showed clear evidence of a multistory nature, indicating that they represent avulsions into pre-existing channels (Fig. 3.12). Mohrig et al. (2000, p. 1799) stated: "The superimposed channel belts within the multistory sandstone bodies in the Guadalope-Matarranya system show clear evidence of reoccupation of a fixed thoroughfare by a succession of channels. The relative frequency of superimposed channel belts suggests that, at least part of the time, abandoned channels remained as relative lows on the flood plain and acted as attractors to avulsing river channels."

Mohrig et al. (2000, p. 1799) noted that "such reoccupation of recently abandoned channels or secondary flood-plain channels is a common characteristic of many modern river avulsions. Examples include the Lower Old River connecting the Mississippi and Atchafalaya Rivers (Fisk 1952), Mississippi River flood-plain channels (Kesel et al. 1974), Rainbow Creek (Brizga and Finlayson 1990), the Assiniboine River flood-plain channels (Rainne 1990), and the River Rapti, in India (Richards et al. 1993)." Pre-existing channels provide ready-made low points and conduits for flow; and unconsolidated sandy channel fills may be easier to scour out than surrounding flood-plain material (Smith et al. 1998). Reoccupation of topographic lows should produce stacked channel sandstone bodies in the stratigraphic record.

It is commonly thought that alluvial ridges, constructed by continuous channel-levee sedimentation along an alluvial plain, would be unlikely to be "reoccupied" by avulsions, because they stand above the floodplain elevation. This is a characteristic built into early alluvial architecture models (Allen 1978; Bridge and Leeder 1979; Mackey and Bridge 1995). Yet reoccupation of channels, as

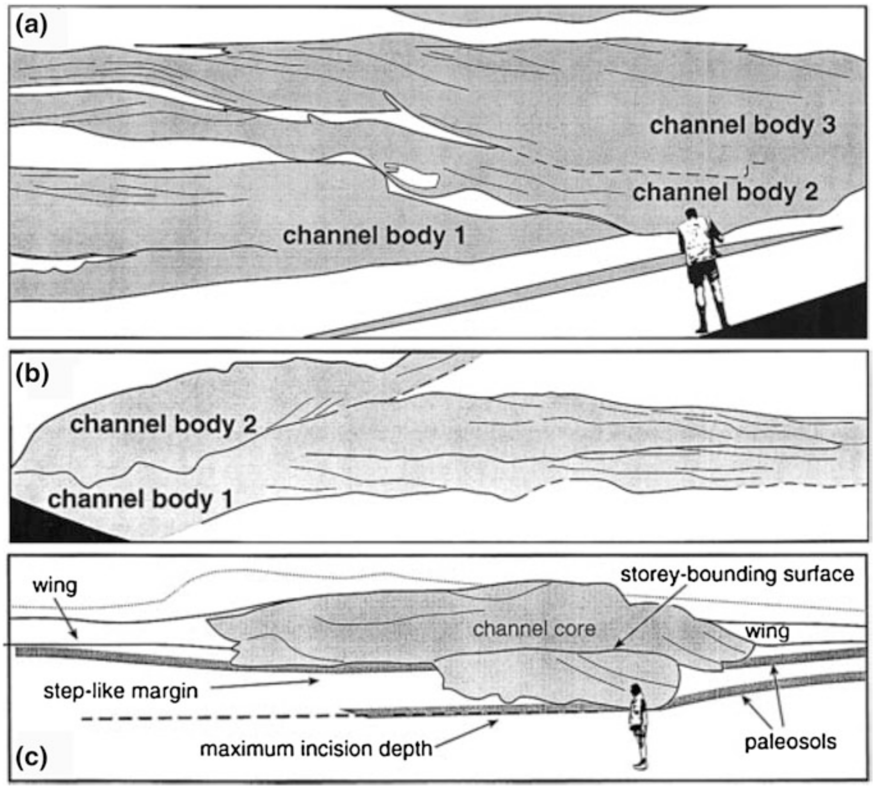


Fig. 3.12 Examples of stacked channel fills in the Guadalope-Matarranya System, Ebro Basin, Spain, which demonstrate a common process, the reoccupation of pre-existing channels as a result of avulsion events (adapted from Mohrig et al. 2000, Fig. 10, p. 1794)

indicated by their multistory nature, appears to be characteristic of the Guadalope-Matarranya System. What might be the explanation of this paradox? It may be, however, that abandoned channel sites that are not completely filled act as semicontinuous thoroughfares that present preferential paths for avulsions. As such, channel position in the flood plain may not be as random as is often assumed in architecture models (Mohrig et al. 2000, p. 1799).

Their studies led Mohrig et al. (2000) to develop two models for avulsion: (a) incisional avulsion, by knickpoint erosion back up a reoccupied abandoned channel, and (b) aggradational avulsion—the overtopping of banks to generate an initial crevasse channel and splay. They suggested (p. 1801): "… that floodplain drainage characteristics exert major control between the incisional and aggradational avulsion modes. Poorly drained flood plains can support standing water bodies that would tend to slow flood waters, forcing deposition. In contrast, well-drained flood plains promote through flowing channel systems and associated early erosion by a variety of mechanisms."

Jones and Hajek (2007, p. 125) stated that:

> Part of the confusion of characterizing avulsion stratigraphy in the ancient is one of defining avulsion. Kraus and Wells (1999) define ancient avulsions by recognizing heterolithic avulsion deposits in a stratigraphic succession. This recognition criterion assumes that all avulsions occur by successive sediment and water diversion from the trunk channel via widespread crevasse splay progradation onto the adjacent floodplain. River avulsion however is more generally defined as an established channel relocating to a new position on the floodplain (e.g., Mohrig et al. 2000; Slingerland and Smith 2004), regardless of the specific processes by which sediment and water are diverted during channel relocation. We propose that this basic definition is an appropriate starting point for interpreting ancient stratigraphy.

Two distinct manifestations of avulsion stratigraphy were observed by Jones and Hajek (2007). Stratigraphically transitional channels are directly preceded by non-overbank deposits, including crevasse splays, crevasse channels, distributary channels, distributary mouth bars, etc. The deposits underlying channel sandstones may contain the suite of "heterolithic avulsion deposits" defined by Kraus and Wells (1999). In contrast, channels juxtaposed directly atop floodplain/overbank deposits are classified as stratigraphically abrupt. There is no evidence of heterolithic avulsion deposits or other non-overbank deposits in proximity to channels within the stratigraphically abrupt successions observed by Jones and Hajek (2007).

The well-exposed Oligocene-Miocene Loranca Basin in east-central Spain has yielded several useful studies of alluvial architecture. The first examined here (Martinius 2000) is a regional study of the large-scale facies architecture, and the nature and origin of architectural change downstream and vertically, through time. The second study (Díaz-Molina and Muñoz-García 2010) takes a detailed look at the internal architecture of meander-belt sandbodies within this system, with a view to reconstructing sand body connectedness and illustrating calculations of reservoir volumes.

The Loranca fluvial system is derived from an active fold-thrust belt, and extends for approximately 60 km downslope in a west to northwesterly direction. Figure 3.13 is a reconstruction of the range of fluvial styles that deposited this fan-like body, and Figs. 3.14 and 3.15 are reconstructed architectural panels that illustrate the depositional architecture at mid-fan and distal fan settings, respectively. The locations of these panels, within the spectrum of fluvial environments, is indicated in Fig. 3.13. The mid-fan location is near the village of Villarejo Seco, and the distal location is near Huete.

The overall environmental interpretation is based on facies and architectural analysis. Conglomerates are a significant component of the system only in the southeast (facies zone 1). "Stacked bar" and "giant bar" sandstone bodies are compared to the linguoid bar macroforms of the modern Platte River (facies zone 2). Ribbon sandstones, with width-depth ratios of between 3 and 25, are comparable to the anastomosed channel deposits described by Smith et al. (1989). Composite point-bar sandstones are the deposits of mixed-load meandering channel systems (facies zone 3). Non-channelized lobate sandstone bodies are interpreted as terminal splay deposits or as crevasse splays (facies zone 4). The variation between these four fluvial styles is summarized in Fig. 3.16. This variation is interpreted as allogenic in origin, because the deposits of each facies zone extend across the full

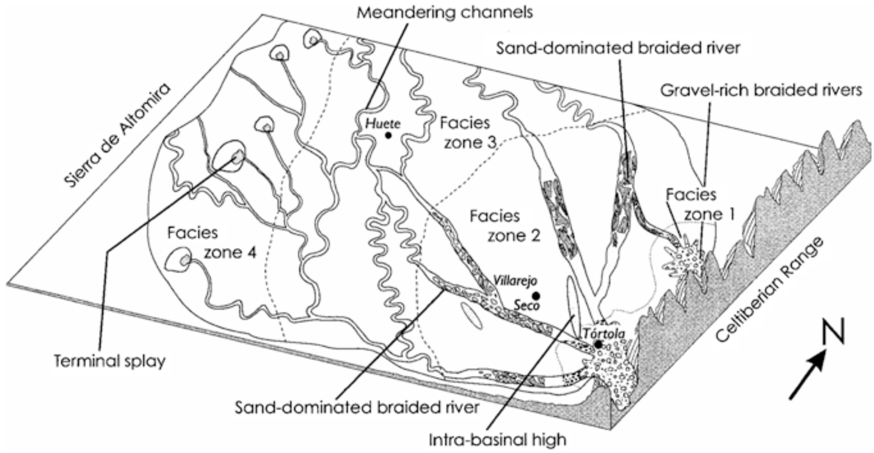


Fig. 3.13 The interpreted fluvial paleogeography of the Loranca fluvial system, Spain (Martinius 2000, Fig. 8, p. 858)

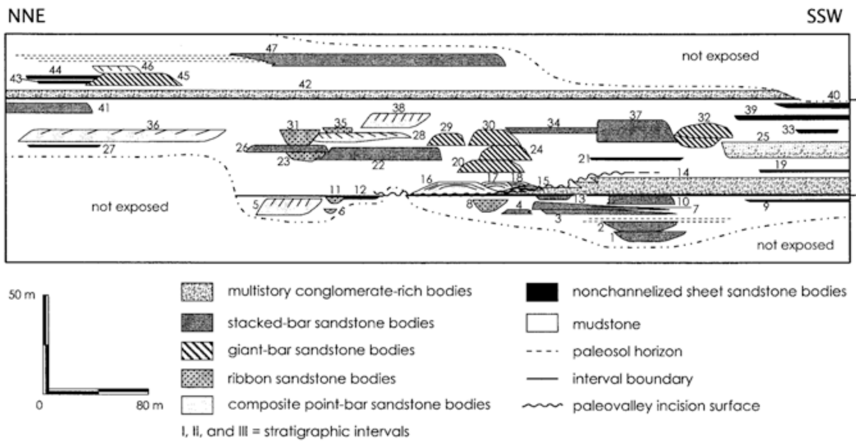


Fig. 3.14 Architectural panel diagram, constructed from exposures in a medial setting in the Loranca fluvial system, near Villarejo Seco (see Fig. 3.13) (Martinius 2000, Fig. 10, p. 860)

width of the outcrop belt, suggesting that the changes in fluvial style are basin-wide in extent. Chronostratigraphic data for the succession suggests that each facies cycle (the change from, say, facies zone 1 to 2, or 3 to 4, in Fig. 3.16), takes about 600 to 800 ka. Although there are subtle indicators of changing climate, none can be clearly correlated with sedimentological changes within this system, and Martinius (2000) concluded that either tectonism or climate change could have been the cause of the changes in fluvial style.

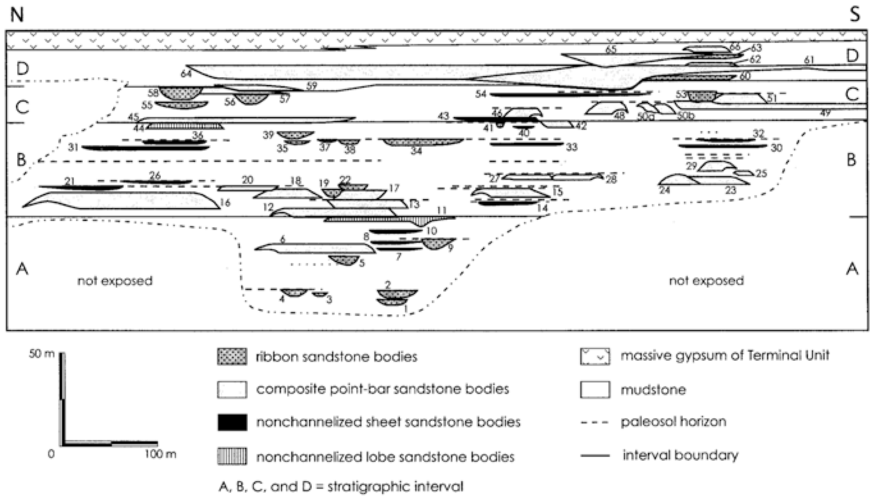


Fig. 3.15 Architectural panel diagram, constructed from exposures in a distal setting in the Loranca fluvial system, near Huete (see Fig. 3.13) (Martinius 2000, Fig. 9. p. 859)

Díaz-Molina and Muñoz-García (2010) focused on the facies architecture of the meander-belt sandstones, near Huete (location in Fig. 3.13), examining them as analogs for reservoir deposits. Figure 3.17a illustrates one of the reconstructed meander-belt sand bodies, in which outcrop observations have been extrapolated back into the near-subsurface, based on the known or most likely architecture of individual meanders. The upper part of most of the point-bar deposits consist of fine-grained mudstones, displaying lateral-accretion bedding. The volume of interconnected sandstone was then defined, based on interpreted sand-on-sand contacts within the section, but excluding the fine-grained upper point-bar deposits from the calculation (Fig. 3.17b). These estimates, when applied to the entire meander-belt succession indicated that sandstone comprises between 38 and 46 % of the total rock volume. Some 92 % of the meander loops were estimated to be vertically or laterally interconnected. These architectural reconstructions also allowed calculations regarding the probability of encountering sandstone in any given well (Fig. 3.18). The probability of encountering at least one meander loop sandstone varied between 36 and 98 %, in the two test areas, while the probability of encountering two such sandstones ranged between 35 and 70 %.

The Huesca fluvial system (Miocene) of the Ebro Basin, in northeast Spain, shows a similar downstream gradation from conglomeratic through braided to meandering sandstone fluvial styles (Hirst 1991; Donselaar and Overeem 2008). Donselaar and Overeem (2008) carried out a detailed study of one of the meander belts of this system (Fig. 3.19), in which sandstone architectures and connectivity are comparable to those of the Loranca meander belts. They suggested two models for the reservoir architecture (Fig. 3.20), one in which the point-bar

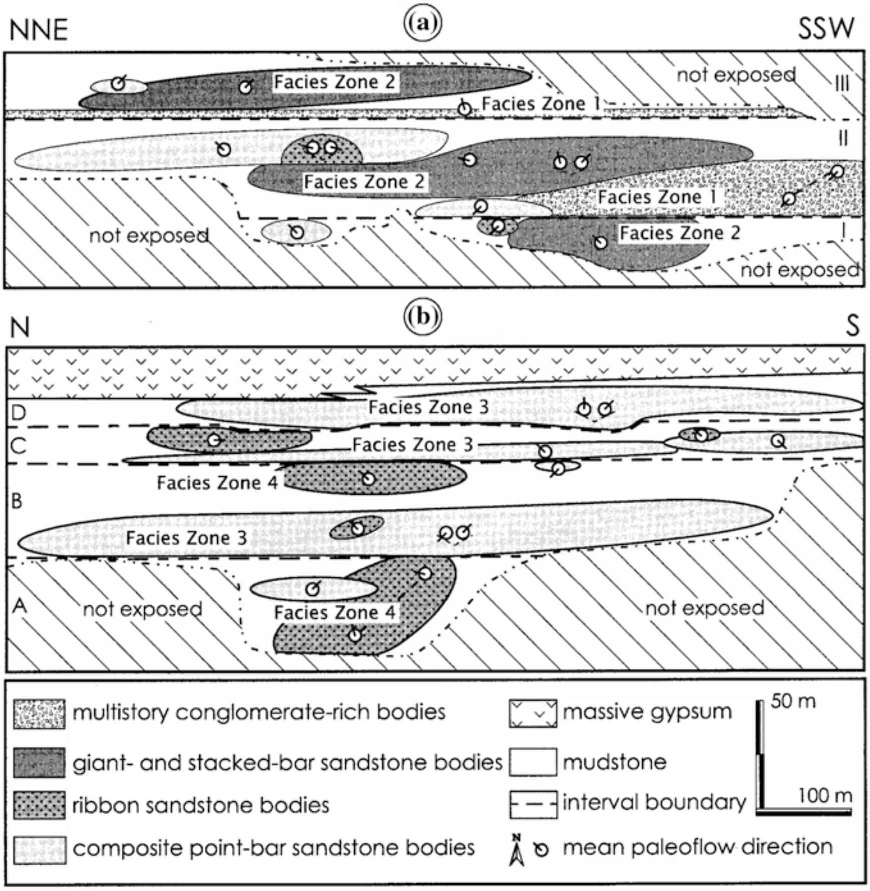


Fig. 3.16 Schematic representation of the cross sections of the (**a**) medial and (**b**) distal areas of the Loranca fluvial system (Figs. 3.14 and 3.15, respectively). Sandstone bodies are grouped according to their position in one of the facies zones of the depositional model. The figure illustrates the stratigraphic position and order of appearance of the characteristic deposits of the facies zones (Martinius 2000, Fig. 13, p. 864)

sandstones are isolated from each other, and one in which they are connected by the relatively coarse, porous sandstones that accumulate on the channel floor. The latter they termed the "string-of-beads" model. This is the preferred or expected model for the Huesca deposits. Estimated sandstone content (or net-to-gross ratio) of the total system thickness is in the order of 40 %, which simulation studies have shown indicates a sandbody connectivity of 90 % (Larue and Hovadik 2006), as discussed in the next section.

A completely different geological setting was the focus of the study by Ryesth et al. (1998). This paper presented the results of a detailed subsurface mapping exercise carried out as the basis of a development plan for the Osberg field in

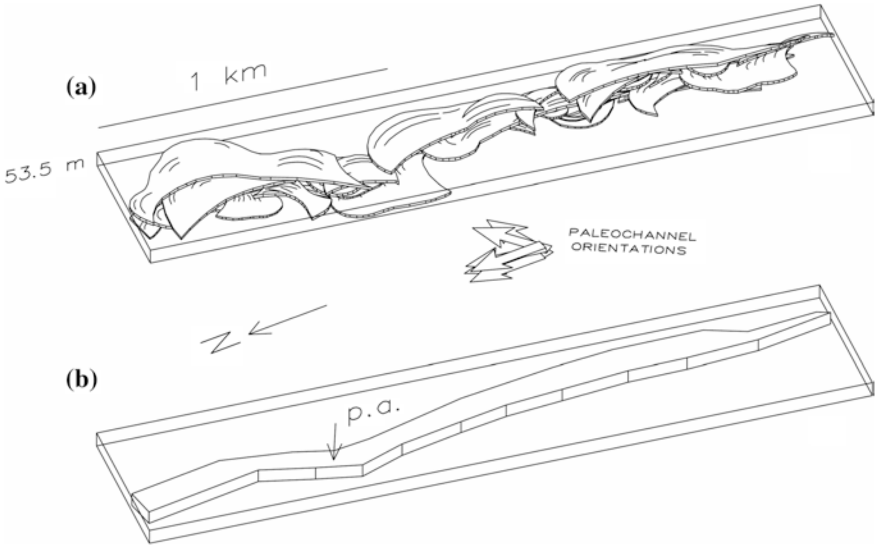


Fig. 3.17 **a** Three-dimensional architecture of seven meander-belt deposits obtained from the compilation of geometric data and paleochannel orientations from the Garcinarro meander belts, Loranca fluvial system. **b** Reservoir volume defined for the purpose of calculation. p.a. = projection area (Díaz-Molina and Muñoz-García 2010, Fig. 7, p. 250). AAPG © 2010. Reprinted by permission of the AAPG whose permission is required for further use

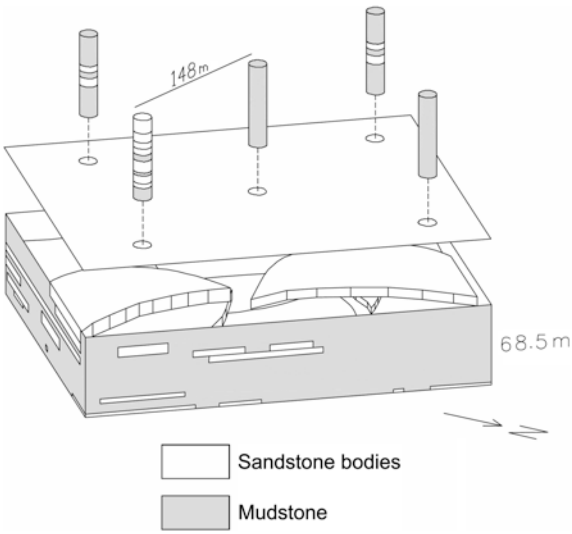


Fig. 3.18 Block diagram with hypothetical drill locations through meander system such as that shown in Fig. 3.17, showing variations in sandstone content over short distances of well separation. Note that some wells will encounter no sandstone at all (Díaz-Molina and Muñoz-García 2010, Fig. 10, p. 253). AAPG © 2010. Reprinted by permission of the AAPG whose permission is required for further use

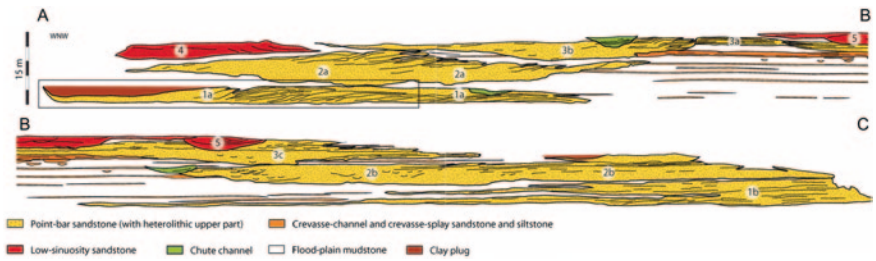


Fig. 3.19 Cross-section through one of the meander belts of the Huesca fluvial fan system of the Ebro Basin, Spain (Donselaar and Overeem 2008, Fig. 5, pp. 1114–1115). AAPG © 2008. Reprinted by permission of the AAPG whose permission is required for further use

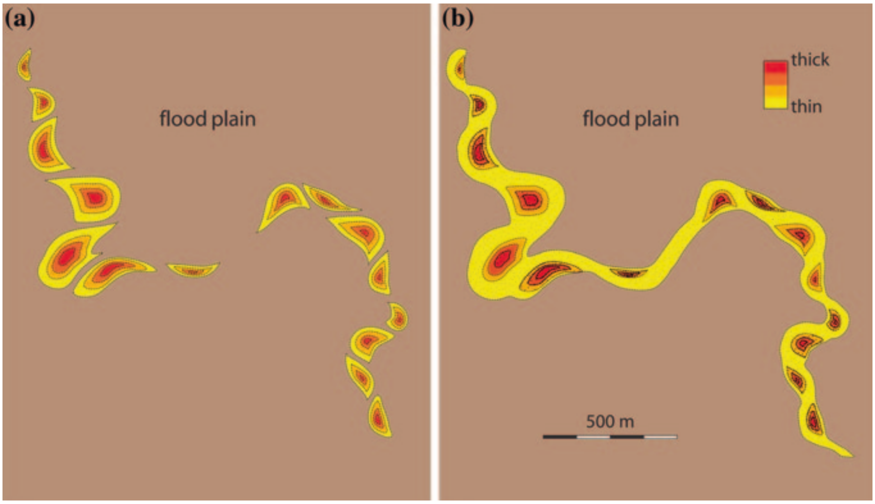


Fig. 3.20 Two conceptual models of the architecture of meander-belt sandstones, based on the Huesca system outcrops (Fig. 3.19). **a** Some point-bar deposits appear to be isolated, which makes for poor reservoir connectivity. **b** In other cases, it appears that points bars are connected by the channel-floor deposits, which consist of relatively coarse, porous sandstone. This is referred to as the "string-of-beads" model of sandstone architecture (Donselaar and Overeem 2008, Fig. 13, p. 1125). AAPG © 2008. Reprinted by permission of the AAPG whose permission is required for further use

the Norwegian sector of the North Sea Basin. This field produces from the Ness Formation, a unit of the very productive Brent Group (Middle Jurassic). Detailed core studies, coupled with wireline log correlation and biostratigraphic control provided the basic control for a sequence-stratigraphic model of sedimentation. The unit consists of an assemblage of channel, overbank, crevasse-splay and related facies representing a complex of meandering fluvial systems (Fig. 3.21).

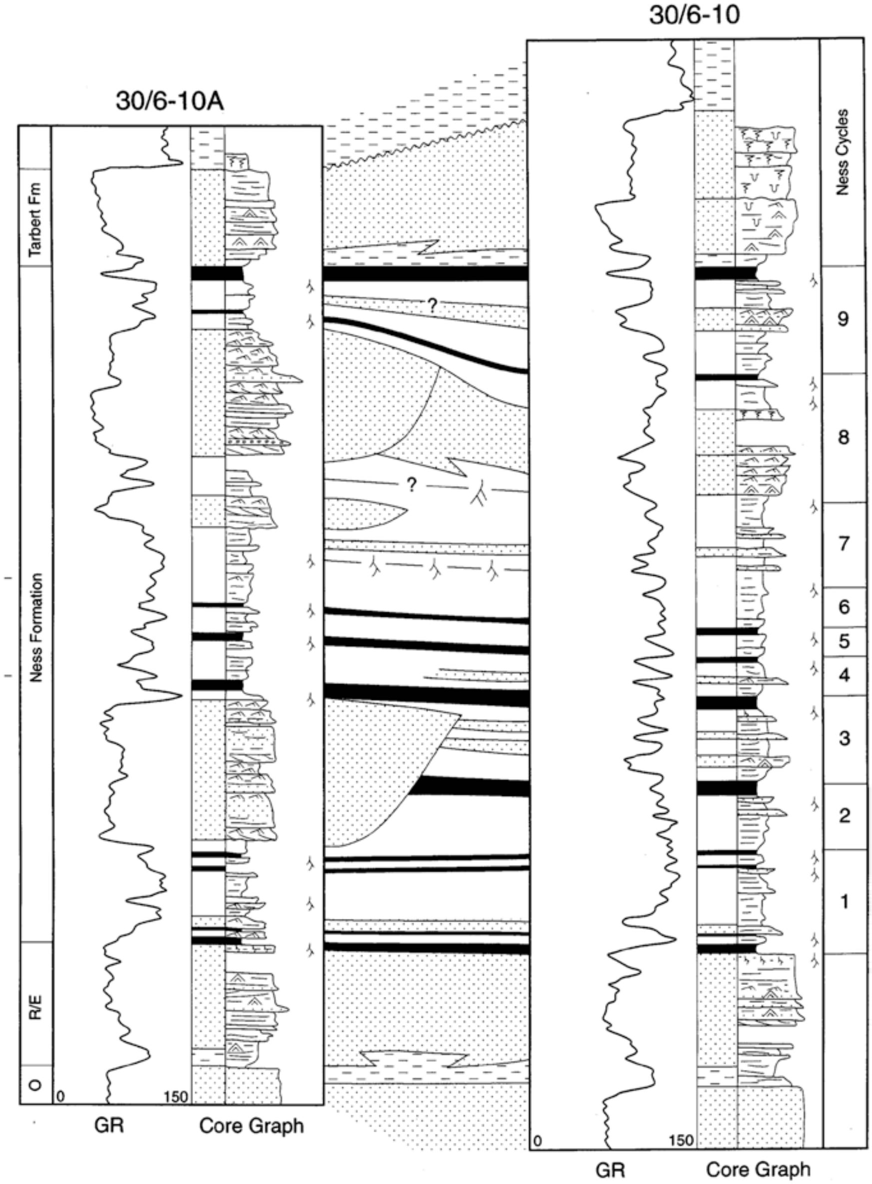


Fig. 3.21 Detailed correlation between two wells 250 m apart in the Osberg Field, North Sea Basin, showing the basin-wide cycles that have been mapped throughout the field (Ryseth et al. 1998, Fig. 7, p. 1636). AAPG © 1998. Reprinted by permission of the AAPG whose permission is required for further use

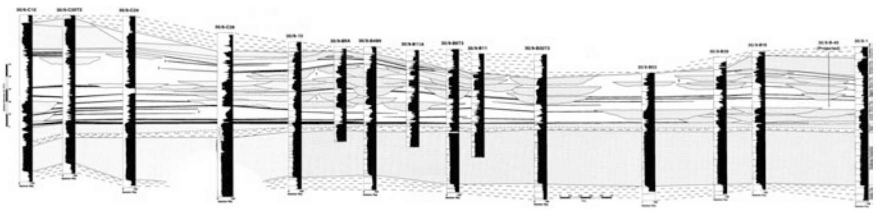


Fig. 3.22 One of a series of correlation panels constructed to show the distribution and interrelation of channel sandstone bodies (Ryseth et al. 1998, Fig. 12, p. 1636). AAPG © 1998. Reprinted by permission of the AAPG whose permission is required for further use

Seismic data were used to plot the likely positions of sandstone-rich volumes, with deviated wells directed to penetrate as many of these zones as possible. We focus here on the facies architecture of these deposits, and return in later chapters to a discussion of the mapping methods and the sequence model.

Figure 3.21 shows the detailed correlation that can be carried out between wells 250 m apart in the Osberg Field. This detailed comparison of vertical profiles suggests that some beds, notably the overbank units (mainly the coals) may be correlated between wells, and this approach has led the authors to subdivide the Ness Formation into 12 “zones” that, it is claimed, may be traced regionally. Correlation panels, such as that shown in Fig. 3.22, do not reveal any systematic architectural patterns, Channels are locally clustered, but are not related in any obvious way to the locations of faults that cut the succession and that, based on thickness variations across the region, were active during sedimentation. Ryseth et al. (1998) interpreted channel distribution as primarily autogenic in origin, a model that does not seem compatible with a model of regional field-wide cyclicity, which would argue for an allogenic control on sedimentation. A sequence interpretation was not suggested by the authors for this succession, apart from the presence of an erosional, partially incised base to the Ness Formation, which is interpreted as an incised valley formed at a time of low accommodation.

Figures 3.23 and 3.24 illustrate the difficulties in attempting to use sedimentological mapping criteria to locate the major features of the paleogeography. A map of channel deposit proportion (CDP) constructed for the Upper Ness Formation suggests a concentration of channels oriented northeasterly and crossing the northern and south-central part of the field area; but a more detailed examination of correlations in one zone of the Ness Formation generated the map shown as Fig. 3.24, which suggests an entirely different alluvial pattern—major channels oriented roughly south-to-north. Isopach maps, sand maps and other contour maps constructed from stratigraphic and sedimentologic data may be very misleading unless they relate strictly to tightly defined units that provide a snapshot of a depositional system over a very limited period of time. We return to this study in Chap. 4 to illustrate the use of subsurface data in the planning of the placement of deviated wells that comprise the production array for the Osberg Field.

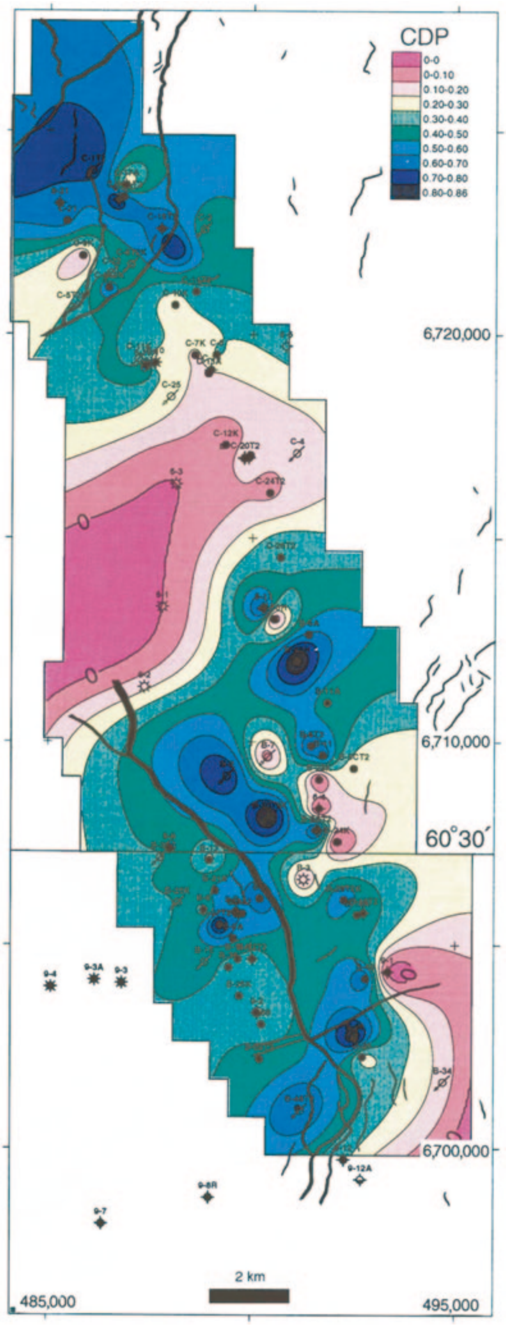


Fig. 3.23 Computer-contoured map of the "channel deposit proportion" (CDP) of the UN1 zone of the Ness Formation, Osberg Field. This zone represents all but the upper, largely fine-grained portion of the Upper Ness Formation. (Ryseth et al. 1998, Fig. 15, p. 1643). AAPG © 1998. Reprinted by permission of the AAPG whose permission is required for further use

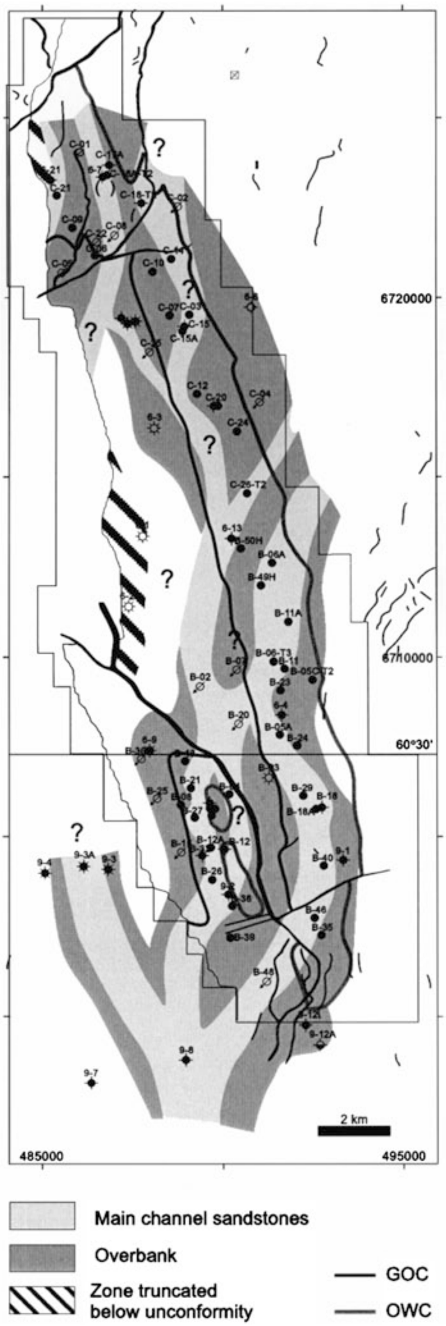


Fig. 3.24 Reconstruction of the main zones of channel and overbank sandstones in zone UN1b of the Ness Formation, one of four zones comprising the Upper Ness Formation. (Ryseth et al. 1998, Fig. 16, p. 1644). AAPG © 1998. Reprinted by permission of the AAPG whose permission is required for further use

A different type of avulsive stratigraphy was described by Prochnow et al. (2006) and Cleveland et al. (2007). These authors explored the nature of fluvial cyclicity in the Triassic Chinle Formation of the Colorado Plateau area, a unit deposited by braided rivers. Cleveland et al. (2007, p. 921) stated “In the Upper Triassic strata of this study, a three-tier hierarchy is present that includes meter-scale fluvial aggradational cycles (FACs) which stack into decametre-scale fluvial aggradational cycle sets (FACSETs) which, in turn, stack into fluvial sequences.” (Fig. 3.25) The FAC and FACSET terminology is based on Atchley et al. (2004). Prochnow et al. (2006) attributed the three-tier hierarchy of FACs, FACSETs, and fluvial sequences in the Late Triassic Western Interior to channel avulsion and floodplain aggradation, successive episodes of channel avulsion, and tectonic changes, respectively. FACSETs were interpreted to have been produced by successive episodes of avulsion as channels drifted away from or toward a given position in the alluvial valley (Kraus 1987; Kraus and Aslan 1999; Atchley et al. 2004; see Sect. 5.2.1).

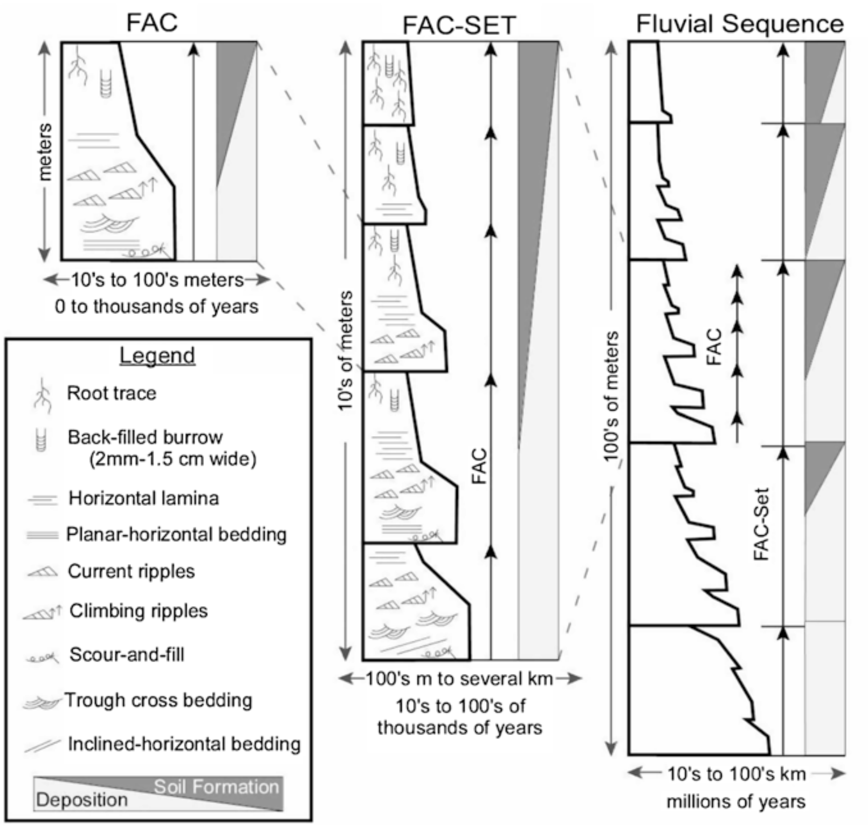


Fig. 3.25 Definition of FAC, FACSET and fluvial sequence in the Chinle Formation (Cleveland et al. 2007, Fig. 7, p. 917; terminology based in Atchley et al. 2004)

A reconstructed stratigraphic cross-section through the Chinle Formation southwest of Moab, Utah, is illustrated in Fig. 3.26. Individual cycles (FACs) are generally fining-upward cycles, capped by paleosols or by discontinuity surfaces. The cycles cannot be correlated between sections. They each represent individual autogenic depositional events within part of the braided fluvial system. Cycle sets may be correlated more widely, as indicated in Fig. 3.25. These consist of stacked FACs which show an overall upward fining and increase in paleosol maturity, suggesting decreasing subsidence rates and exposure time through the accumulation of each FACSET. Prochnow et al. (2006) attribute the development of FACSET succession to periodic subsidence due to salt withdrawal in the underlying Paleozoic section, a topic to which we return in Chap. 5.

On a basin scale, channel systems may form *channel clusters*, when viewed in cross-section, and the locations of these clusters may provide insights into large scale (allogenic) depositional controls. Good examples of this were described by Hajek et al. (2010) and Hofmann et al. (2011). As the latter documented from a careful subsurface study, channel bodies were organized stratigraphically into clusters, and these clusters became systematically displaced, or offset through time (Fig. 3.27). Hajek et al. (2010) used statistical methods to reject the hypothesis of

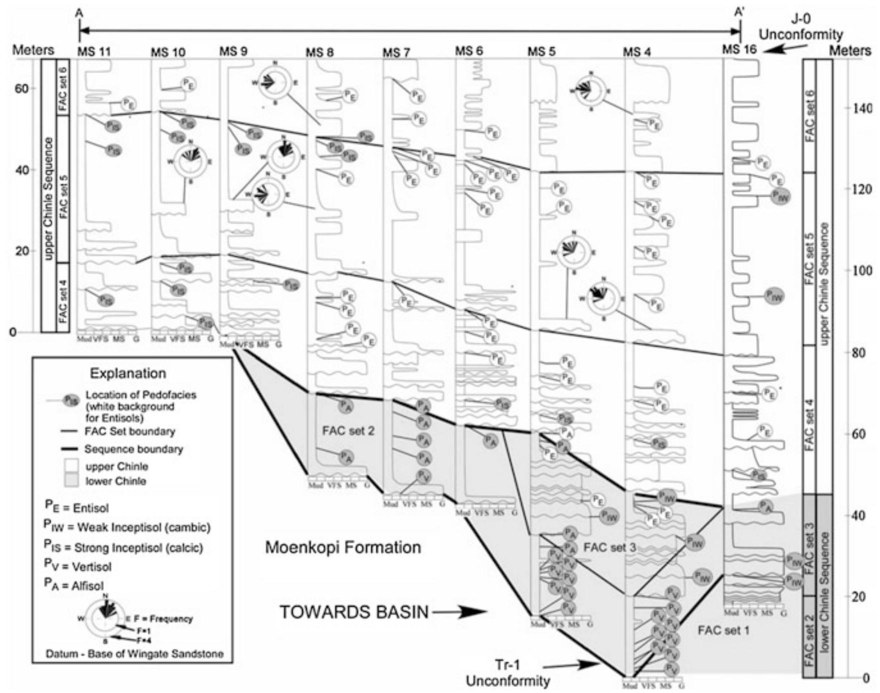


Fig. 3.26 Cross-section through the Chinle formation, near Moab, Utah. The section is approximately 12 km in length, and shows the succession of fluvial aggradational cycles (FACs) and cycle sets (FACSETs) in this formation (Prochnow et al. 2006, Fig. 5, p. 1333)

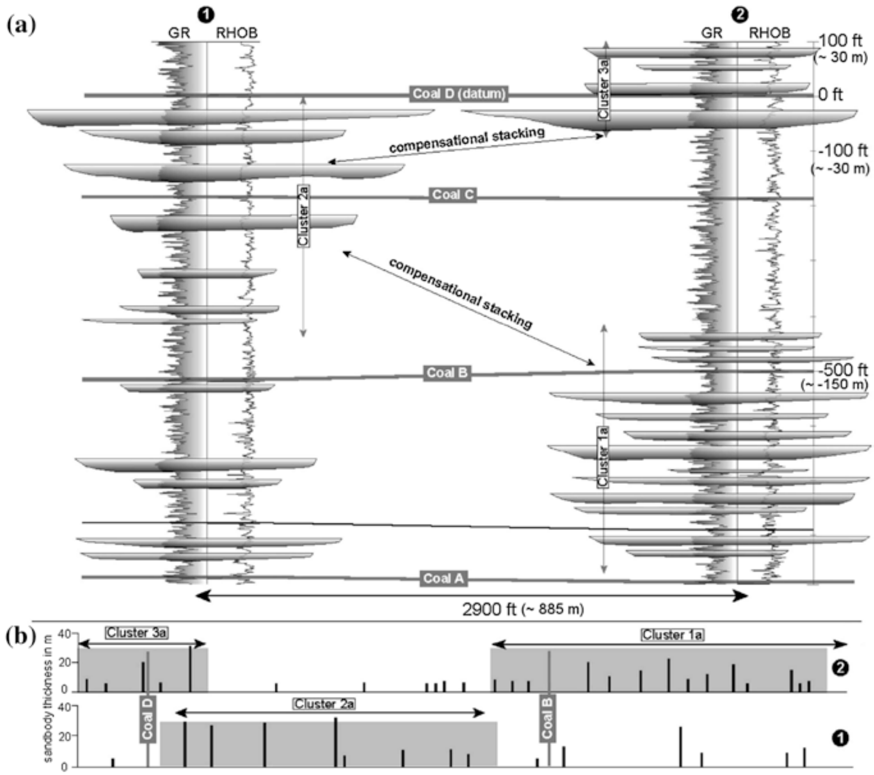


Fig. 3.27 Channel clusters and *compensational stacking* in the Williams Fork Formation (Upper Cretaceous), Piceance Basin, Colorado (Hofmann et al. 2011, Fig. 3, p. 674)

random channel stacking. These patterns suggest that clusters may preferentially fill available accommodation, leading to lateral displacement, much as delta and submarine-fan lobes are displaced at the front of these depositional systems, a process termed *compensational stacking*, a process we examine briefly in a later section.

The processes whereby channel clusters formed in the case of the Williams Fork Formation (Hofmann et al. 2011) are the autogenic processes discussed in this section. The localization of the channels into clusters was interpreted by Hofmann et al. (2011) as having been determined by the location of alluvial "fairways," which may have represented individual major (trunk) rivers in the system. In many of the clusters it was observed that, on average, channel bodies became thicker upward within each cluster, suggesting an increasing tendency to channel amalgamation. Most clusters are capped by a widespread coal.

Hofmann et al. (2011) interpreted the formation of clusters with upward increases in sandbody amalgamation as a response to decreasing accommodation rates (but see Sect. 6.2). Compensational stacking may represent the effects of an allogenic control, causing long-term changes in accommodation over an estimated 400 ka time periodicity.

3.5 Numerical and Scale Modeling of Avulsion

Modern attempts to understand the development of alluvial architecture began with the speculations of Allen (1965a) regarding the expected stratigraphy of the main types of river, including braided, meandering, and low-sinuosity single-thread rivers. Allen's later (1974) detailed study of the Devonian Old Red Sandstone in the Welsh borderland area has come to constitute a major milestone in the understanding of the controls on fluvial architecture. This unit contains numerous pedogenic carbonate horizons that Allen realized represent major pause surfaces on the alluvial plain, and that could be employed as time markers to correlate preserved channel systems. The various patterns of channel and floodplain relationship present in the Old Red Sandstone led Allen to speculate regarding the possible autogenic and allogenic controls in sedimentation. Autogenic processes included the "combing" of channels across a fan-like distributary plain, as had been observed in the Kosi fan of India (Gole and Chitale 1966), and avulsive switching. Allogenic processes included climate change and base-level change. Figure 3.28 illustrates his model of avulsive stratigraphy. Another of his models constitutes the first realistic model of base-level control of alluvial architecture, and is in effect the first sequence model for fluvial systems. We illustrate this in Chap. 6.

Numerical models have now been widely employed to simulate fluvial processes with the objective of bringing greater understanding and predictability to studies of the ancient record, particularly the record in the subsurface, where it

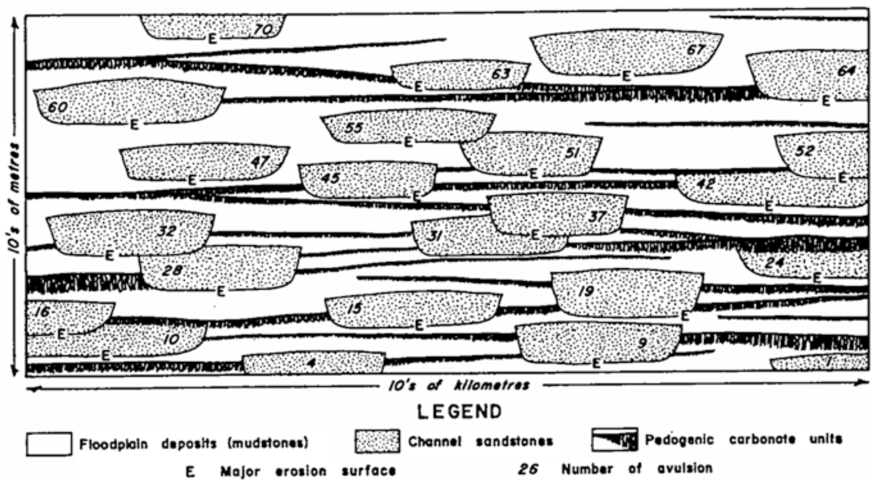


Fig. 3.28 One of the alluvial architecture models of Allen (1974, Fig. 7, p. 197). This is model 3, which illustrates the predicted stratigraphy of an alluvial suite developed by periodic channel avulsion under conditions of steady subsidence

may be of considerable interest to petroleum, exploration geologists and production engineers (see Figs. 2.8 and 2.9 and discussion thereof). All these models originated with the speculations of Allen (1974), which were followed by some simple simulation experiments (Allen 1978, 1979), in which Allen explored the issue of sand body connectivity. In the first of these studies, Allen (1978) imagined a simple, linear, subsiding continental margin, much like the Cenozoic margin of the Gulf of Mexico (Fig. 3.29). In this model (Allen 1978, p. 130–131):

> Differential subsidence is taking place with the consequence that the plain and an imaginary hinge line parallel with the coastline are slowly moving seaward. Along any spatially fixed section AB parallel with the coastline and seaward of the hinge, a uniform and constant rate of subsidence, *R*, is considered to prevail in the short term. The model will show how a sequence of alluvium is built up as the plain subsides beneath AB, the products of its operation being a vertical two-dimensional section of the strata, normal to the transport direction. We finally assume that the entire region is uniform in geological composition and structure, and also in physiography and climate. Rivers of a uniform character may then be regarded crossing the plain at a constant longitudinal spacing. Each river is considered to have a zone of influence of width *W* on the plain, within which it alone is responsible for the channel sand-bodies accumulated.

Allen 1978 (p. 146) concluded that "the character of an alluvial suite may be described in terms of the number-density and area-density of sand bodies, where the latter is equivalent in a statistically isotropic suite to the average sand/mud ratio afforded by one-dimensional vertical profiles (boreholes, conventional logs)." Turning to a property of particular interest to petroleum geologists, he

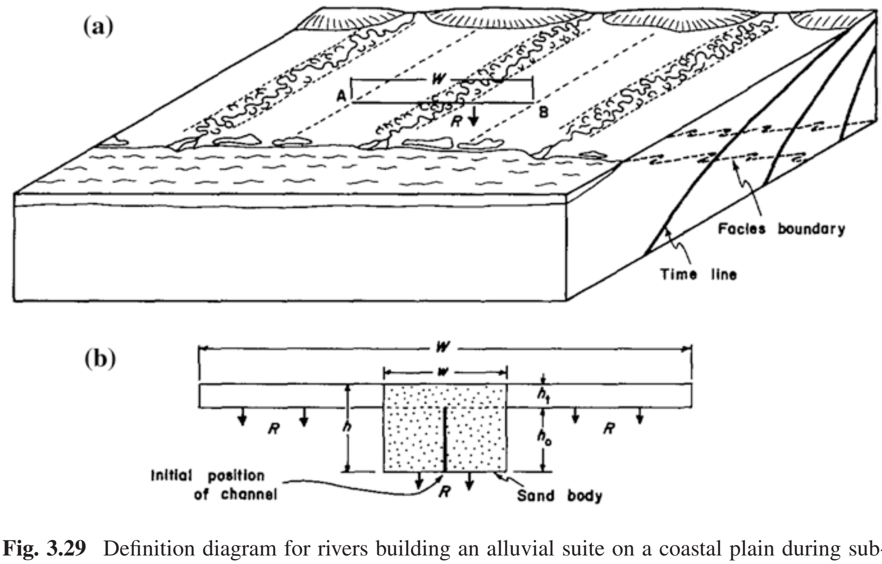


Fig. 3.29 Definition diagram for rivers building an alluvial suite on a coastal plain during subsidence-controlled offlap. **a** General character of plain and zones of river influence. **b** Details of sediment increment constructed between one avulsion and the next (fine-grained overbank sediments shown blank) (Allen 1978, Fig. 1, p. 131)

also demonstrated that "The connectedness of the sand bodies may be described by the coordination (average sand-body contacts per body), domain size (average number of bodies in bounded cluster of bodies), and the average fractional contact (fraction of sand-body perimeter in contact with adjacent bodies)." For sand-body contents (net-to-gross) of less than 50 %, he concluded that the degree of connectedness is very small, whereas above a content of 50 %, the connectedness increases very rapidly.

The most important controlling factor in autogenic models of alluvial stratigraphy is the determination of the rate and style of avulsion. As noted earlier in this chapter, little is known about the detailed avulsion history of most rivers, and so some generalizing assumptions have to be made. Leeder (1978) and Bridge and Leeder (1979), in their models, assumed that avulsion is not dependent on the existing position of the channel within the model, but occurs sufficiently far upstream from the location of the model that the channel is free to move to the current lowest position on the floodplain. Avulsion frequency was determined by reference to the limited information on stratigraphic and historical data, and the recurrence interval was assumed to follow a distribution function similar to that of earthquakes, which in any location of tectonic stress become more likely to occur as time passes. Mean avulsion periods ranging from 111 to 1780 years were used by Bridge and Leeder (1979).

Given an avulsion event the channel is moved to the lowest point on the floodplain, and allowed to accumulate a channel sand body at a preset accumulation rate, ranging from 5 to 40 m/ka. Floodplain accretion is accomplished using a simple equation that reduces the aggradation rate across the width of the model cross-section in proportion to the distance from the channel belt. This reflects the fact that floodplain sediments are derived primarily from overbank flooding. Compaction of the accumulated deposits is accomplished, layer by layer, using preset compaction factors. This process generates relief on the floodplain surface because floodplain fines compact less than channel sands, and the least compaction occurs where channel sands are stacked vertically. The model therefore builds in a feedback effect, in that the site of the next lowest position for channel occupancy will be determined by the outcome of the preceeding sedimentation-compaction history. An example of Bridge and Leeder's (1979) models is illustrated here (Fig. 3.30). In this model, sandbody distribution appears to be relatively random after the initial ten avulsion events. The Bridge and Leeder (1979) model has been much used, particularly by petroleum geologists, because of the insights it yields into channel stacking patterns and interconnectedness, factors of considerable importance in the understanding of reservoir predictability and fluid migration behaviour. However, as discussed in Sects. 6.1 and 6.2, there is a significant issue concerning rates of accommodation generation and sedimentation rates that must be taken into account when applying these models to the rock record.

There is now a fairly long history of such model experimentation, the complexity of which reflects increases in sophistication of the understanding of fluvial processes, the increasing data base of relevant information on which such models are based and the increasing computer power that has facilitated an improvement from

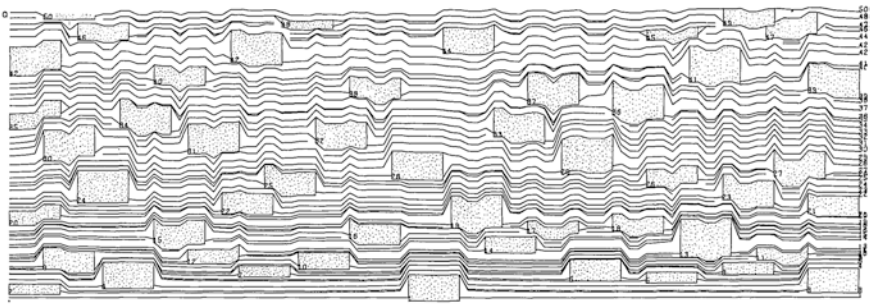


Fig. 3.30 Example of a simulated cross-section of alluvial stratigraphy constructed using the numerical model of Bridge and Leeder (1979, Fig. 2, p. 630). The cross-section is 10 km wide. Channel-belt aggradation rate is 20 m/ka. Channel width is 600 m and depth is 3 m

two- to three-dimensional models and the use of ever more sophisticated graphical presentations of the results. It is interesting, however, to note the statement of Larue and Hovadik (2006, p. 295) that: "It is comforting to realize that most of the conclusions about 2-D connectivity presented by Allen (1978, 1979), using the crudest of modeling techniques, are still embraced today."

Bryant et al. (1995) referred to the founding models of Leeder, Allen and Bridge as the "LAB models" (Allen 1978, 1979; Leeder 1978; Bridge and Leeder 1979). Bryant et al. (1995, p. 365) stated:

> The theme of the LAB model is that as channel belts avulse (switch paths abruptly) in a depositional river system, they produce a distribution of channel-belt sands in the vertical plane ("alluvial architecture") that depends on, among other things, the geometry and rate of deposition and the dynamics of avulsion. A simple and widely used outcome of the LAB model is that, if the frequency of avulsion is constant, the cross-sectional density of buried channel belts should be inversely correlated with sedimentation rate: widely dispersed, isolated "ribbon" channel-belt sand bodies represent high sedimentation rates, when a substantial accumulation of overbank fines is deposited between avulsion events. More densely connected, coalescing sandbodies represent periods of low sedimentation rate, when little overbank material is deposited and channel systems rework the flood plain, redepositing fines downstream. Other possible influences on alluvial architecture in the LAB model include sediment compaction, tectonic tilting, and variation in width of the channel belt (Bridge and Mackey 1993a, c).

Essentially all these models, from Bridge and Leeder (1979) to Mackey and Bridge (1995), are geometric models. They make no attempt to simulate the actual physics of sediment transport, the turbulence of flowing water, the erosion of channel margins, the partitioning of the water and sediment load into a crevasse and out onto a crevasse splay, the compaction of the resulting sedimentary accumulations, and the subsequent iteration of all these processes. Actual digital simulation of the physics on the ground would, of course, be a formidable task absorbing an immense amount of computer power, so, as with all numerical models that are used in the earth sciences, the modeling process makes use of numerical short-cuts that simulate the actual physics. This is not even "parameterization" as it is usually understood. This term refers to the estimation of the range of values to be used as input in a particular set of calculations of some physical process. The numerical models of fluvial avulsion and architecture construction simply make use of geometrical approximations of the actual physics. For example, compaction is an important process in the construction of alluvial architectures because it controls the accommodation that is generated on an alluvial surface as channel bodies and floodplain deposits are accumulated on it during successive avulsions. The physics of compaction involves the importance of the local petrological mixture of clay minerals, silt and other particles, the expulsion of pore waters under increasing hydrostatic load, the ability of the water to escape as controlled by the porosity and permeability of the sediment body (flow equations), considerations of surface tension, the gradual orientation of clay and other particles, and so on. None of this is utilized in the compaction calculations of alluvial architecture models (or those of basin subsidence models, for that matter). Instead, an empirical exponential equation is derived from the best fit of a curve drawn through empirical measurements of porosity changes with depth in selected sampled stratigraphic sections of known lithology.

A major controlling factor in alluvial architecture models is the elevation of the channel bankfull flow relative to that of the floodplain. Figure 3.2 illustrates the typical structure of the *alluvial ridge*, with its raised levees. This referred to as the *superelevation* of the channel. Figure 3.31 illustrates two diagrams that provide a definition of this term. The mainstream models of Leeder, Allen and Bridge, and their successors, focus much attention on these elevation differences across the simulated alluvial plain, because they are employed within the model to provide

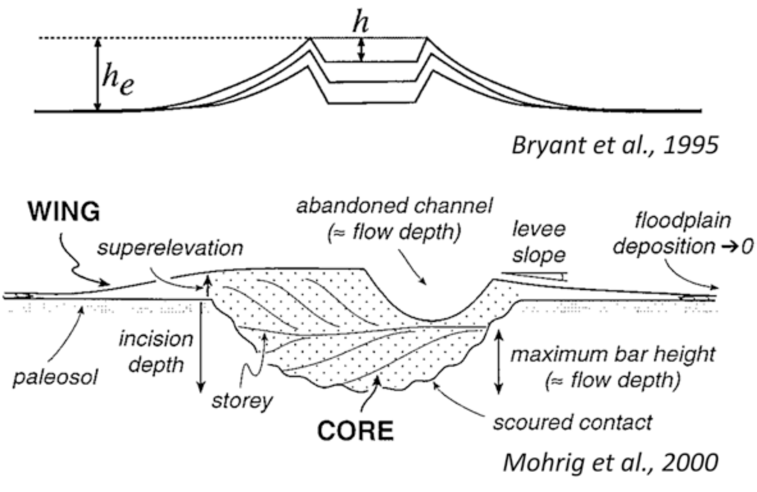


Fig. 3.31 Two illustrations of the key parameter controlling avulsion, the superelevation of the channel above the floodplain (h_e in the top diagram). A "wing" is the preserved levee deposit, the wedge of bedload deposited on the flanks of the channel

the triggers for diversion and the direction of flow when diversion takes place. However, it is all geometry, based, admittedly on the actual tendency of water to flow down a slope, but paying no attention to the physics of flow around a bend, the erosive power of the river under various conditions to generate a crevasse, the local control of turbulence patterns to facilitate or inhibit erosion and flow diversion, and so on. The actual moment of avulsion is an event simulated by a process of randomization within the model, based on the geometry of the valley and channel slopes and of channel and floodplain elevations simulated from input sedimentation-rate values. Stochastic devices are employed, making use of what is termed Monte Carlo methods—the repeated random sampling from a range of potential input values exhibiting a predetermined probability distribution. Nature is not, of course, random, but commonly the processes are so complicated that they appear random, and this is why Monte Carlo methods have become so widely employed. But it's not physics.

Mackey and Bridge (1995, and building on earlier studies) were able to generate some crude simulations of such distinctive architectural processes as the gradual lateral shifting of the major distributary of the Kosi River, India. However, the simulation was far from perfect, and though the model looked elegant, the question arises, so what? How does this assist us with the interpretation and prediction required for future exploration and production work? Some generalities emerged about the role of tectonism, the construction of alluvial ridges along the floodplain, and so on, but none of these was surprising. It is important, too, to bear in mind the issues of accommodation rates and sedimentation rates, discussed in Chap. 6.

A modeling study of avulsion was reported by Bryant et al. (1995), who stated (p. 366): "The connection between avulsion and channel superelevation resulting from preferential deposition near the channel axis suggests that avulsion rate could be related to sedimentation rate, if differential sedimentation between channels and flood plain is scaled to overall sedimentation rate." They measured avulsion frequency as a function of sediment-feed rate in a series of small scale model fluvial fans. Their experiments demonstrated a strong relationship between avulsion frequency and sedimentation rate. It remains unclear how the relationship might change over time, as the model fan was gradually aggraded.

Slingerland and Smith (1998) examined the standard model of avulsion—the building of an alluvial ridge higher than the floodplain. They approached the problem by employing the equations for the physics of sediment transport to explore what might happen once an initial crevasse had been eroded by overbank flooding. They noted (p. 435): "Our results show that whether a crevasse of arbitrary initial geometry will trigger an avulsion depends upon sediment grain size, the initial depth of the crevasse, and the ratio of crevasse to main channel bed slopes." Reoccupation of a pre-existing channel, a common occurrence, as discussed below, will tend to provide a deeper crevasse.

In a series of papers (Ashworth et al. 1999, 2004, 2007) examined the avulsion characteristics of gravel braided rivers and compared model data with field date from exposures of the Recent gravel braided river deposits of the Canterbury Plains, South Island New Zealand. In general, the results were as might have been expected, avulsion frequency is positively related to sediment supply, as influenced, in the natural setting, by tectonic and/or climatic factors.

Oreskes et al. (1994) provided an insightful critique of numerical models as employed in the earth sciences. Numerical models are widely employed, for example, to predict groundwater flow, reservoir productivity, climate change, to name three examples. Numerical modeling is used in alluvial architecture studies because of the assumed value of such models to assist in the prediction of reservoir architectures for future drilling programs. However, two quotes from Oreskes et al. (1994, p. 643) provide food for thought.

> If the predicted distribution of dependent data in a numerical model matches observational data, either in the field or laboratory, then the modeler may be tempted to claim that the model was verified. To do so would be to commit a logical fallacy, the fallacy of "affirming the consequent."
>
> … even if a model result is consistent with present and past observational data, there is no guarantee that the model will perform at an equal level when used to predict the future. First, there may be small errors in input data that do not impact the fit of the model under the time frame for which historical data are available, but which, when extrapolated over much larger time frames, do generate significant deviations. Second, a match between model results and present observations is no guarantee that future conditions will be similar, because natural systems are dynamic and may change in unanticipated ways.

The conclusion seems to be that if the results of numerical models are useful in clarifying concepts, or, in the case of reservoir simulation, provide useful preliminary results, then their use may be continued, but extrapolation too far beyond initial conditions in time or space should be carried out with caution. As discussed in Chap. 2, fluvial style is rarely constant in directions either parallel or perpendicular to structural grain in sedimentary basins, and numerical simulation therefore has obvious limits.

A different but related question is: how realistic are the results of small-scale laboratory experiments? Can they really be "scaled up" to provide insights into actual fluvial processes? In a wide-ranging review, Paola et al. (2009) argued that many natural geomorphic processes are to a considerable extent scale independent, which therefore justifies the use of small-scale experiments to model natural processes. The discovery of the fractal character of the elements of many natural systems provides powerful support for this assertion. In an earlier paper by this research group, Sheets et al. (2002) argued that scale modeling was ideally suited to an investigation of what they termed mesoscale dynamics, that is, the intermediate time scale (10^1–10^4 years) between the formation of channels and bars on the observable human time scale, and the long time scale "on which autocyclic variability sums to produce the average behaviour represented in large-scale stratigraphic models" and which, in natural systems, begins to approach the precision of high-resolution stratigraphic dating techniques (10^5–10^6 years). In a series of runs, they explored the predictions of the LAB avulsion models. Laboratory experimentation essentially simulates the braiding process. It has not been found possible to develop models that successfully mimic meandering rivers. Avulsion has a different meaning in this setting. Whereas in meandering and anastomosed systems avulsion represents a shift of most or all of the discharge from a major channel into an entirely different setting within the floodplain, in braided rivers, avulsion refers to the shift of individual channels amongst a spectrum of large and small channels within what is usually a wide braid plain.

Sheets et al. (2002) observed two contrasting types of avulsion event. In some cases, flooding overtopping banks led to reoccupation of earlier channels, and the consequent diversion of a portion of the braided channel flow into a new (earlier) location. In other cases, where there was no pre-existing channel to reoccupy, a broad crevasse splay sheet was deposited. Channel shifting was accomplished primarily by avulsion rather than migration. Short-lived avulsions were found to be most effective in filling the topography and therefore developing a potentially preservable stratigraphic record, whereas "successful" avulsions served as sediment transport routes, with no net deposition. There was also found to be an inverse relationship between channel flow occupation time and net aggradation. Long-lived channels served as sediment transport routes, with little net deposition, whereas channels that were abandoned accumulated a significant aggradational fill as abandonment took place.

On a short-term basis (the mesoscale), the pattern of avulsion and deposition does not demonstrate a response to subsidence or tilt patterns scaled to model natural rates. In other words, allogenic processes on these scales do not appear to influence autogenic processes. However, given enough time, some consistencies emerged. Sheets et al. (2002) demonstrated that in their model, after the deposition of a sediment layer equivalent to between five and ten channel-depths, which required an equivalent number of "successful" avulsion events to occur, the resultant layer had evolved a relatively consistent thickness and that the regional variation in this thickness could be related to the pattern of subsidence. These observations suggest a means to relate subsidence rates to avulsion frequencies, and together they offer significant insights into the formation of the stratigraphic record.

Jerolmack and Mohrig (2007) explored a simple relationship for avulsion frequency, f_A, using field and laboratory data. Their equation $f_A = \nu_A N/h_m$, where ν_A = representative (near channel) aggradation rate, N = number of active channels, h_m = average channel depth, proved to be a useful descriptor of average river behaviour, and was not related directly to specific triggering events, such as floods or ice jams. They suggested that anastomosis of individual rivers, and the development of deltaic distributaries may arise from the same process, the main difference being that rivers tend to be confined between valley walls, whereas deltas are typically unconfined coastal systems.

A later experimental study, by Straub et al. (2009) carried these observations further. This project was focused on the mechanisms behind the formation of *compensational stacking*. Compensational stacking is an architecture in which delta lobes and submarine fan lobes are successively offset laterally from one another, a process long interpreted as the result of slope advantages diverting discharge and sediment into low areas in a depositional system (Fig. 3.32). Given enough time, these depositional lobes may return to the same map position, resulting in offset

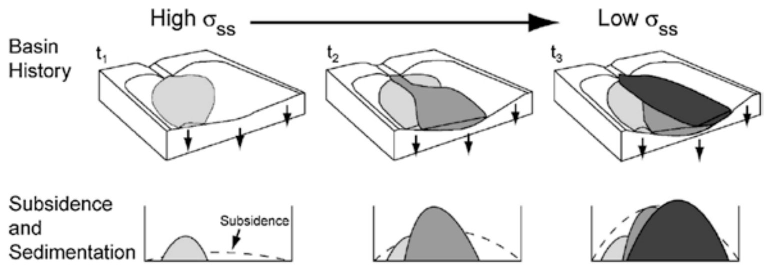


Fig. 3.32 The process by which deposition is gradually distributed across an entire distributive basin, such as an alluvial fan, delta or submarine fan. In deltas the lateral displacement is typically driven by slope advantages, triggered by an avulsion event. In fluvial systems this process is recognizable by the clustering of channels and the rapid displacement of clusters to different parts of the basin. Sedimentation is rapid where clusters are forming, but is slow, zero or negative (erosional) elsewhere in the system, but over time, the average overall sedimentation rate will average that of subsidence rate (*dashed line*). This is formalized by the expression σ_{ss} = standard deviation of sedimentation/subsidence. Successive clusters form following avulsion into the unfilled accommodation indicated by the *white spaces* below the *dashed line*. This can lead to *compensational stacking* of clusters (and delta/submarine fan) lobes. Diagram from Straub et al. (2009, Fig. 2, p. 676)

stacking through time. Similar architectures may occur in fluvial systems, as documented in a study by Hofmann et al. (2011) to which we make reference above (see Fig. 3.27). Straub et al. (2009) carried out a series of experiments and performed a detailed examination of a seismic data set from the Mississippi delta to explore the natural controls on compensational stacking. They devised a measure, κ, the compensation index. For pure compensational stacking, i.e., when deposition always fills topographic lows, $\kappa = 1.0$. Where stacking is random, without any influence of pre-existing topography, $\kappa = 0.5$. Anticompensation, where low areas become lower as a result of in-place stacking, $\kappa = 0$. Their experiments and observations indicated that natural systems typically may have values of κ between 0.5 and 1.0, but values in the mid-range between these extremes were typical. This result is of interest and potential value in the construction of reservoir models.

Wang et al. (2011) defined a compensational time scale, $T_c = \iota/r_a$, where ι is the surface roughness, and r_a is the average, long-term sedimentation rate. Roughness, in this setting, refers to the topographic mounding that occurs as a result of avulsion and the building of alluvial ridges. They determined that the time scale relationship "essentially states that the geometry of deposits carries the signature of stochastic autogenic dynamics out to a time scale equal to the time necessary to fill a basin to a depth equal to the amount of surface roughness in a transport system. It is interesting that T_c for many systems extends into time scales commonly associated with large-scale allogenic cycles (e.g., Milankovitch cycles)" (Wang et al. 2011, p. 814). This means that purely autogenic processes may take much longer than had been assumed to work though in an alluvial basin, and that this may take such processes into the time scales normally assumed for allogenic controls (10^4–10^5 years and more).

From the point of view of this book, one of the more important observations made by Paola's group, modeling fluvial systems in their experimental facility, is that "unambiguous chronostratigraphic horizons are rare. Indeed, study of experimental fluvial strata highlights how elusive 'time surfaces' really are in highly dynamic sedimentary systems. The experimental deposit is just an accumulation of fragmentary surfaces that cannot be precisely correlated over distances of more than (at best) a few channel widths." (Sheets et al. 2002, p. 294). This has implication for sequence stratigraphy, as we return to in Chap. 6.

3.6 Sand Body Connectivity

In an important study based on a review of oil fields in Texas, Tyler et al. (1984) and Tyler and Finley (1991) argued that oil recovery is strongly dependent on facies architecture, and that improvements in our understanding of, and knowledge of, the architecture of complex reservoir successions could substantially increase oil recovery factors, by providing improved guidance for the placement of production wells, and for the design of secondary and tertiary recovery projects (see Fig. 2.10, which is from these studies). Studies of this type have been a major impetus for sedimentary geologists in a program of research to improve our understanding of channel geometry (based on fluvial style), and stacking patterns (based on autogenic and allogenic controls on the development of alluvial belts). In Sect. 2.1.1 we introduced the problem of interpreting fluvial architecture from limited outcrop and subsurface data. We reviewed the sedimentological fixation with "fluvial style" as a predictor of architecture, and concluded that it is only one factor to consider, and may not even be the most important one. In this chapter we have reviewed the processes of autogenic avulsion as a control on fluvial architecture, and demonstrated the wide range of time scales and avulsive architectures that have been preserved. Reservoir performance is in part based on channel (sand body) connectivity, but this is a very variable and elusive characteristic of fluvial deposits (see Figs. 2.14 to 2.16 and the discussion thereof).

Larue and Friedmann (2005, p. 131–132) pointed out that:

> There is a largely unrecognized controversy regarding the degree to which stratigraphic architecture influences recovery efficiency. As discussed below, geologists tend to believe that stratigraphic architecture and geometric shapes influence recovery strongly, based on connectivity and continuity arguments. Because of this, there has been a tendency in the geological community to make more and more complex and realistic facies models. Engineers, while acknowledging the importance of connectivity and continuity, are more prone to stress permeability heterogeneity and permeability anisotropy on recovery, as well as dynamic reservoir characteristics such as fluid types and well architecture.

Much depends on the range of variability of porosity and permeability in the reservoir sandstones; for example, overall recovery may be significantly reduced by the presence of a local "thief zone" of high permeability, that leads to rapid water breakthrough. Oil viscosity is also a critical factor, influencing the rate at which enhanced recovery may be optimized. These criteria are the domain of production engineers, and have been considered systematically elsewhere (e.g., Thakur 1991; Thakur and Satter 1998).

In this section we review some important modeling (simulation) experiments carried out by Larue and Friedmann (2005) and Larue and Hovadik (2006). In the first of these papers they ran simulations of "channelized reservoirs" deposited in unspecified sedimentary environments, using statistical methods to construct a range of simulated architectures, and then running waterflood programs through each model. They tested three net-to-gross settings, 35 %, 60 % and 85 %, varying channel-sand thickness, channel sinuosity and stacking pattern for each sand percentage. Well spacing (110-acre spacing), water saturation, fluid properties and relative permeabilities were held constant for each mode. Figure 3.33 shows the variation in recovery efficiency for the 60 % net-to-gross experiments, where other parameters were modified one at a time. Figure 3.34 illustrates how each of these parameters affects recovery, when varied one by one, while keeping the other parameters unchanged.

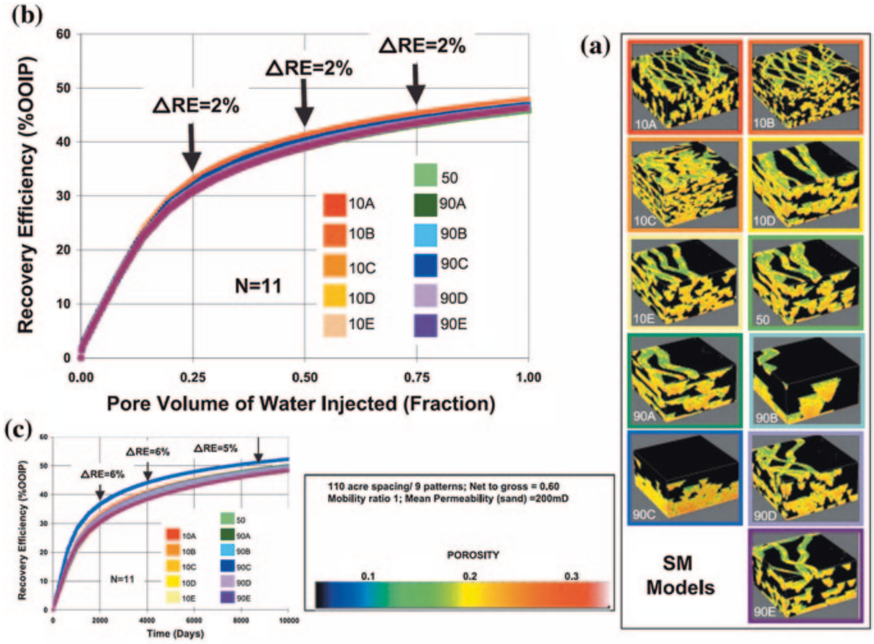


Fig. 3.33 Recovery efficiency in a simulated fluvial architecture with a net-to-gross set at 60 %. The thumbnail sketches illustrated at (**a**) images of porosity for eleven different models where sand body width, thickness, sinuosity and stacking pattern were each varied one at a time while keeping other parameters constant. The spread of recovery efficiencies shown in (**b**) indicates that the variations in these parameters leads to only a 2 % variation in recovery factor, where plotted against injected volumes, but increases to 6 % where plotted against time (**c**) (Larue and Friedmann, 2005, Fig. 6, p. 137)

Not a great deal of variability was apparent. As Larue and Freidmann (2005, p. 137) noted "Even though 33 models with wide range of net: gross were considered here, the total spread in recovery efficiency at 0.5 PVI is only about 5 %. All the stratigraphic uncertainty represented by the wide range of geometries in the models did not have disastrous effects on recovery efficiency."

Of interest is the breakdown in the range of effects generated by each parameter in turn. This is shown in Fig. 3.34. Not surprisingly, the recovery efficiency values increase from the TM to the SM to the EM models, with a variation of up to 4 % depending on how some of the architectural details vary within each of these net: gross model suites. The lower efficiencies corresponding to higher input values (the ranges highlighted with stars in Fig. 3.34) are not explained in detail by Larue and Friedmann (2005), and may relate to factors not directly recorded, such as permeability heterogeneity. Other models run as part of this suite of experiments indicate that this parameter is one of the most important overall influences on reservoir performance, the presence of high-permeable thief zones being particularly troublesome in terms of sweep efficiency.

A companion paper by Larue and Hovadik (2006) focused specifically on channelized reservoirs. They addressed in some detail the relationship between net:gross and connectivity. Allen's (1978) experiments had concluded that connectivity drops off rapidly at net:gross values of less than 50 %. Larue and Hovadik (2006) cited later studies based on percolation theory, which concluded that 66 % is the key ratio. Connectivity is very low until the net:gross reaches this value, and then increases rapidly. Larue and Hovadik (2006) ran several hundred different model simulations in two dimensions, and again in three dimensions,

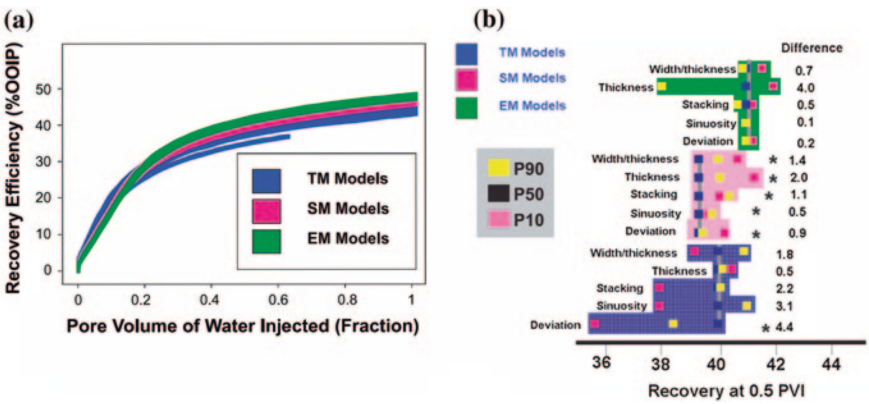


Fig. 3.34 Comparison of recovery efficiency results for three models of net:gross with three sets of input values for sand body width, thickness, stacking, sinuosity and directional deviation. Net: gross models are TM = 30 %, SM = 60 %, EM = 85 %. The range of input values is indicated by the values P10, P50 and P90, for the smallest, median and highest values in each range. Values indicated with a star are "out of sequence", that is, the progression of points from P10 to P50 to P90 is inconsistent. See text for discussion (Larue and Friedmann 2005, Fig. 8, p. 138)

using a wide range of channel configurations in order to test this proposition. The results are summarized in Fig. 3.35. The “S” curve representing the best fit to the data points, is characteristic of all the experiments of this type. The near-vertical portion is referred to as the “cascade zone.” In the 2-D experiments, this zone spans the 50–75 % net:gross ratio. For values more than 75 %, connectivity is greater than 95 %; for net: gross ratios less than 50 % connectivity is effectively zero in this study. For the three-dimensional models the cascade zone spans the 10–30 % range. Given that actual reservoirs are, of course, three-dimensional, this second diagram is of practical relevance, indicating that a minimum net:gross for good reservoir performance is 30 %.

These model results are, of course, based on the randomized sampling and placement of idealized channelized reservoir bodies. Some of the geological realities that may affect reservoir compartmentalization and recovery for better or worse are illustrated in Figs. 3.36 and 3.37. Firstly, of course, it is misleading to focus solely on the channelized units. Crevasse splay sands may provide important additional elements of the sand body connectivity (the yellow units shown in Fig. 3.36a). They may have much broader lateral extents than the channels, and may, therefore dramatically improve connectivity and overall reservoir performance. Channels with large width:depth ratios (Fig. 3.36b), sandstone sheets, such as those generated by flood events, and channel clusters generated at times of low accommodation (Figs. 3.36c , d), are all likely to be characterized by better connectivity. The architecture illustrated in Fig. 3.36d is of particular interest, because it corresponds to the predicted changes in accommodation generation associated with the development of nonmarine sequences. The peak of low rate of floodplain aggradation indicated in this figure is the type of scenario expected to be associated with the generation of a sequence boundary.

Conversely, there are characteristics of natural systems that may intervene to reduce connectivity and reservoir performance. The orientations (paleoflow

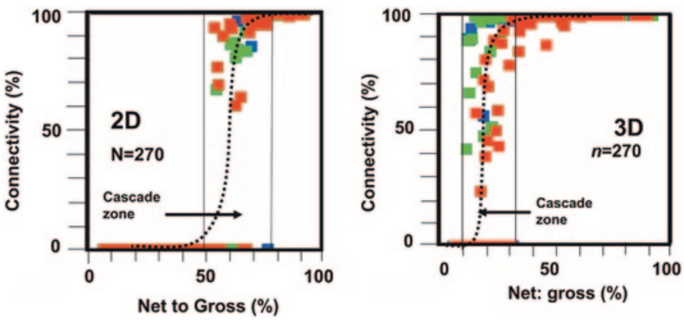


Fig. 3.35 Connectivity as a function of net:gross, in a suite of two-dimensional experiments (*left*) and three-dimensional experiments (*right*). The different colours refer to different channel configurations, the details of which are not significant for our purposes (Larue and Hovadik 2006, Fig. 6, p. 297; Fig. 7, p. 298)

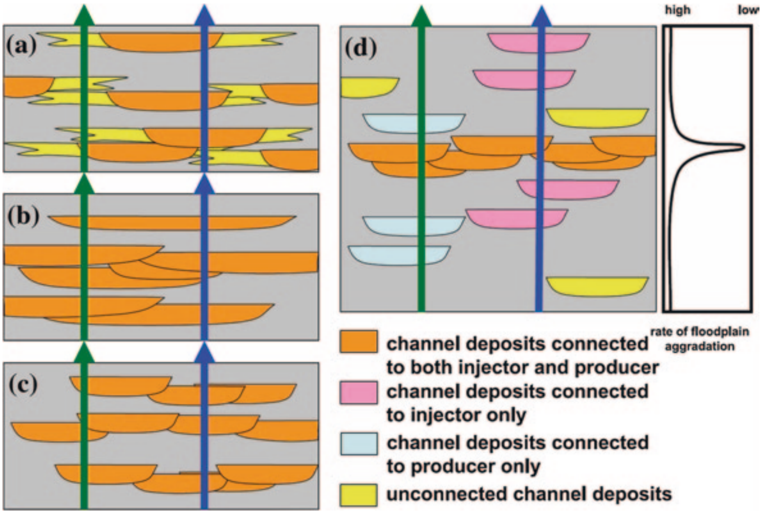


Fig. 3.36 A range of channel stacking scenarios that enhance sand body connectivity. See text for discussion (Larue and Hovadik 2006, Fig. 10, p. 301)

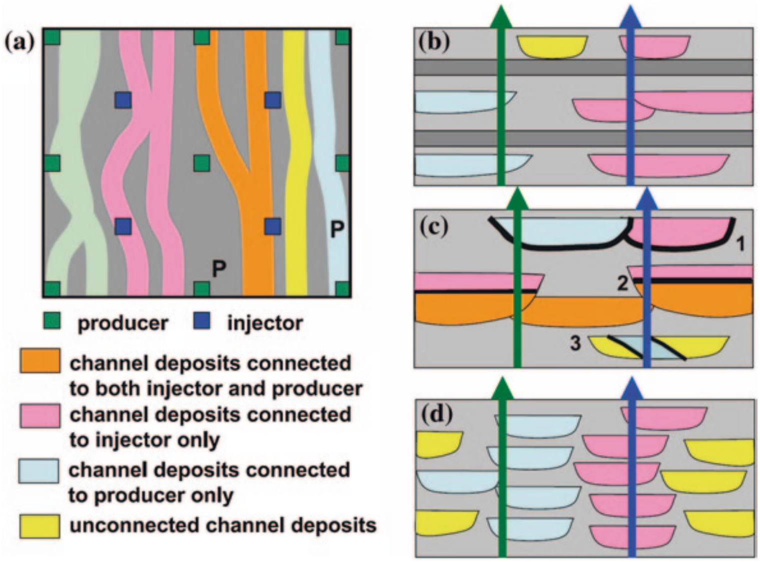


Fig. 3.37 A range of channel stacking scenarios that reduce sand body connectivity. See text for discussion (Larue and Hovadik 2006, Fig. 11, p. 301)

directions) of contemporaneous and successive fluvial systems may have developed geometries that reduce connectivity, for example parallel channels for which no cross-channel connections are present (Fig. 3.37a). The converse of the low-accommodation, high-connectivity situation shown in Fig. 3.36d is that illustrated in Fig. 3.37b. Here, widespread regional mudstone intervals impede vertical connectivity. This is a facies association to be expected during high-accommodation periods in basin evolution—that correlated with the transgressive systems tract in coastal, marine-influenced settings.

Within-channel facies variations may be significant. These are illustrated schematically in Fig. 3.37c, and include mudstone drapes and plugs (1), stratigraphic compartmentalization generated where the channel is composed of multiple successive fills, some or all capped by mudstone (2), and channel fills developed by successive lateral-accretion events separated by inclined, low-flow mudstone units (3). The latter situation is particularly characteristic of channel systems composed of inclined heterolithic units (Thomas et al. 1987). Reijenstein et al. (2011) illustrate examples of channel systems that contain well-developed point bar complexes composed of porous sand, cross-cut by mud-filled channels "potentially isolating one point bar from another, creating reservoir compartments" (op. cit., p. 1986).

The final scenario, that shown in Fig. 3.37d, is described as "compensational stacking" of channels. This may potentially arise where channel deposition builds a ridge on the basin floor (such as the alluvial ridge shown in Fig. 3.2), so that as subsidence and accumulation continues, there is a tendency for subsequent deposition to be diverted towards lower areas within the basin. This results in the offsetting of channels over the long term. Compensational stacking of channels and depositional lobes has been observed in some submarine fan systems (e.g., the Frigg fan in the North Sea basin: Hertier et al. 1980; see also examples cited by Larue and Hovadik 2006, p. 303). In the case of fluvial systems (some examples are noted earlier in this chapter, e.g., see Fig. 3.27), diversion by avulsion away from the alluvial ridge into the flood basin is the expected course of events, a discussion of which is the major topic of this chapter.

An example of a practical application of these concepts was described by Pranter et al. (2009). The Lower Williams Fork Formation, in the Piceance Basin, Colorado, is a gas producer. It outcrops near Grand Junction, Colorado, where the stratigraphic architecture was studied in detail for the insights it might provide on production strategies. Detailed vertical sections were measured through the unit along a well-exposed outcrop belt some 9 km in length. The widths of exposed channel bodies were estimated, taking their orientation relative to the outcrop into account by using paleocurrent data. The vertical sections were then treated as if they were subsurface well sections, with subsurface simulation models then developed using appropriately spaced sections and the statistics of the channel widths, depths and connectivities to construct production scenarios with 10-acre and 20-acre spacings (Fig. 3.38). The exercise indicated that production could be increased by infill drilling.

Labourdette (2011) used a well-exposed channel complex in Spain as the basis for a series of simulation experiments to explore reservoir-body connectivity (Fig. 3.39). The channels are part of a system of eight sequences, which are interpreted as the product of high-frequency tectonism. Each sequence is characterized

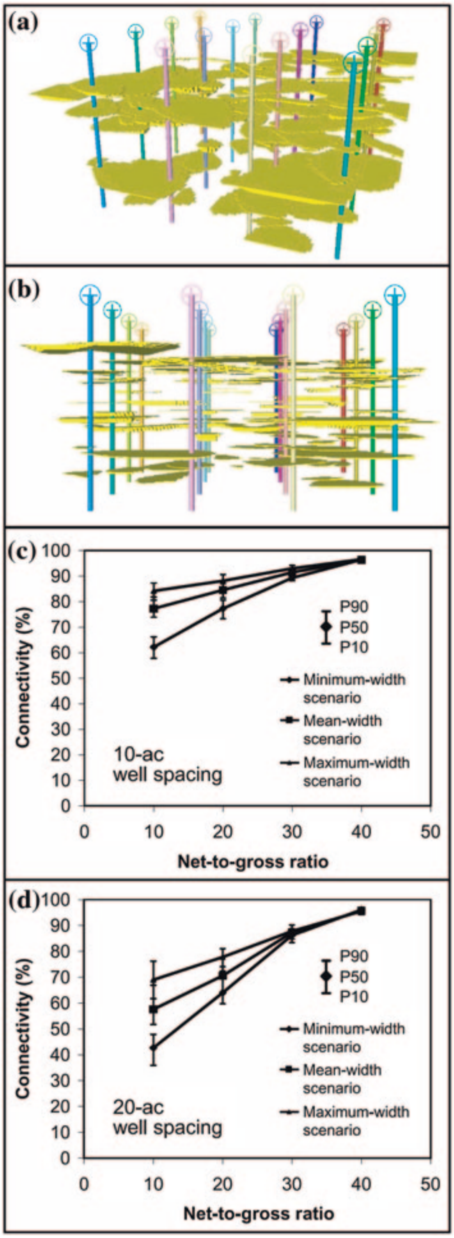


Fig. 3.38 **a**, **b**: Two views of a 3-D architectural model for the Williams Fork Formation of the Piceance Basin, constructed using minimum widths for the channels, a 10 % net:gross ratio and a 10-acre well spacing. **c** and **d**: Are graphs of net:gross ratio versus sandstone-body connectivity for three width scenarios at 10- and 20-acre spacings, respectively. The P10, P50, and P90 scenarios indicate minimum, median and maximum values (Pranter et al. 2009, Fig. 15, p. 1398). AAPG © 2009. Reprinted by permission of the AAPG whose permission is required for further use

by a set of "low-accommodation" channels at the base, consisting of multistory, laterally-amalgamated sandstone bodies, and a "high-accommodation" system at the top, in which the channels are narrower and thicker (Fig. 3.40). No attempt was made in this paper to explore in detail the controls on channel architecture.

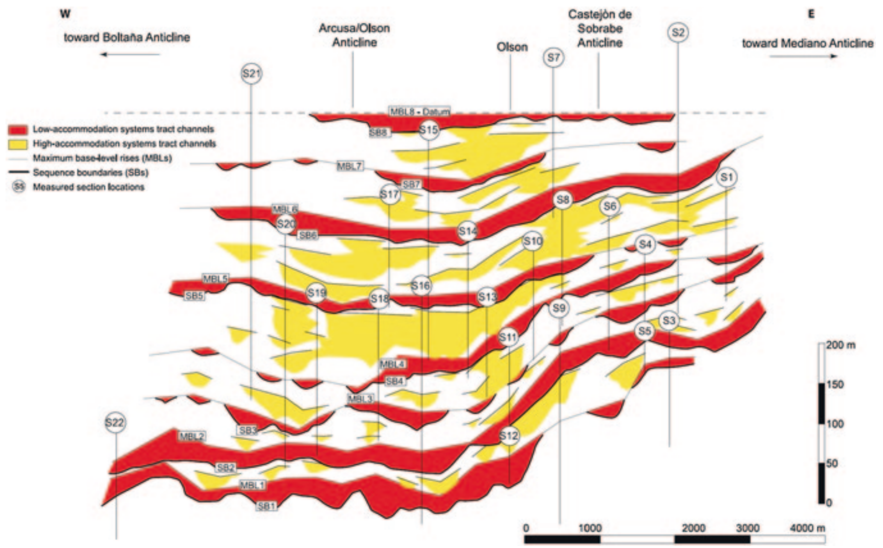


Fig. 3.39 Outcrop cross section of the Lower Olsen member of the Escanilla Formation, based on outcrop measurements and aerial photograph interpretation (Labourdette 2011, Fig. 17, 608). AAPG © 2009. Reprinted by permission of the AAPG whose permission is required for further use

The simulation model was run using a model size of 4000 × 3000 × 200 m, with a cell size of 15 × 15 × 1 m. The dimensional data from the outcrop channels was used to construct a size distribution, which was then used as the basis for sampling in the construction of four successive modeling experiments (Fig. 3.41). Other facies, including floodplain deposits, were not considered to have any potential to add to sand body connectivity. For the first model runs, channel sand bodies as a proportion of the total volume were kept to the actual average of 40 %, but the stratigraphic distribution was random, paying no attention to distribution within the sequences. The result is a simulation that contains 31 reservoir bodies with connected volumes ranging between 1336 and 6.68×10^8 m^3.

In the second model, simulation was constrained by the stratigraphic framework. This was done by developing a "vertical proportion curve" to constrain the vertical distribution of channel bodies. This imposed a distribution such that low-accommodation intervals were simulated at 80 % channel proportion, and high-accommodation intervals at 20 % channel proportion. The simulation consisted of 26 connected reservoir bodies, with a connected volume ranging between 169 and 5.85×10^8 m^3. Most of the connected volume consists of wide, multilateral sheet sandstones in the simulated low-accommodation intervals.

In the third model, additional constraints were imposed on the model to reflect subtle variations in differential subsidence across the field area. It had been observed that the section thins slightly over two intraformational anticlines that cross the area (Fig. 3.39), and that the low-accommodation channels were slightly

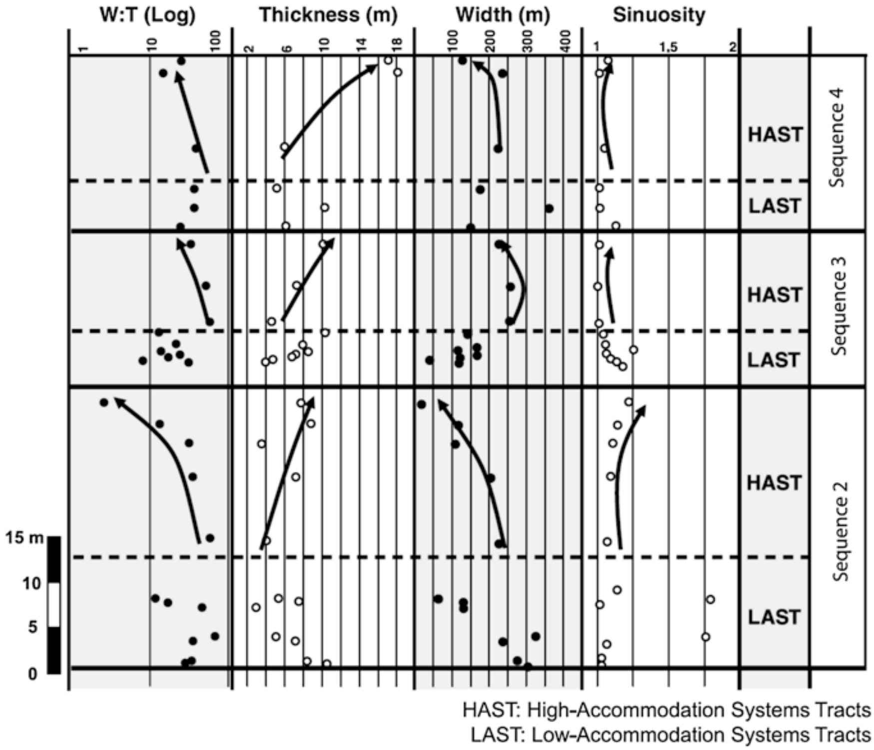


Fig. 3.40 Vertical changes in channel morphology through the Lower Olsen Member of the Escanilla Formation (Labourdette 2011, Fig. 15, p. 604). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

thinner in these areas. The depositional topography imposed by this intraformational tectonism had an effect on contemporary paleoflow, and resulted in high-accommodation channels being preferentially concentrated in the synclinal areas. The result was an increased number of connected reservoir bodies (46 discrete bodies), with a range of connected volumes of between 203 and 8.66×10^8 m^3.

In the final model, channel volumes were constrained by vertical changes in morphology. Field work had indicated that there were no such changes in the low-accommodation intervals, whereas in the high-accommodation settings field observations had indicated that within each interval there was a decrease in channel width and an increase in channel thickness with height (Fig. 3.40). This resulted in a further increase in the simulated connected volume. The simulated volume of connected geobodies ranged between 139 and 1.246×10^9 m^3.

The lesson learned from these simulations is that local stratigraphy matters. The volume of simulated connected sandstone bodies increased with each addition of a geological constraint. In addition, the focusing of the main connected bodies in the synclinal areas means that recognition and mapping of these subtle

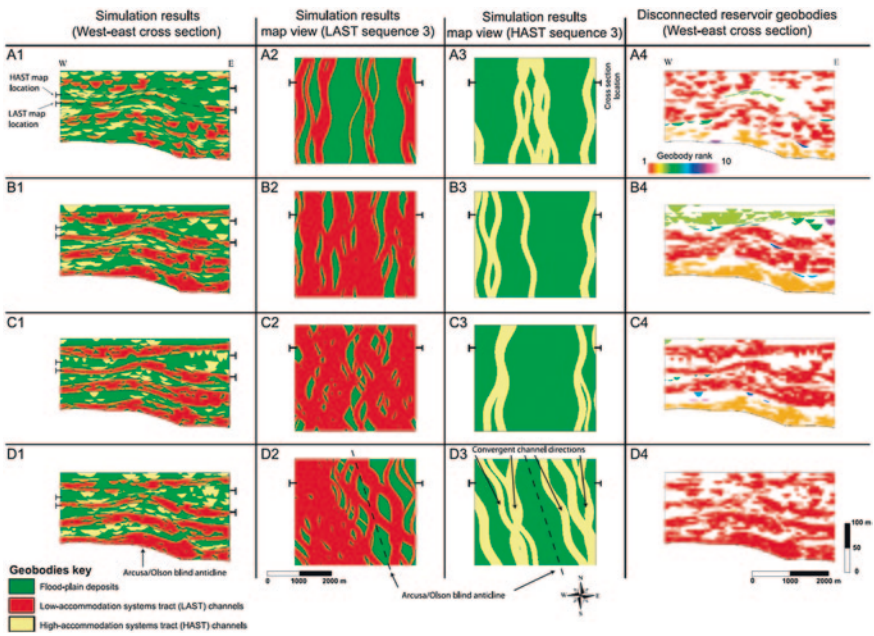


Fig. 3.41 Four simulations of the channel distribution in the Lower Olsen Member of the Escanilla Formation, showing simulated cross-sesctions (*column 1*), maps (*columns 2 and 3*) and geobody ranking (*coumns 4*). Rows A to D refer to the four models discussed in the text (Labourdette 2011, Fig. 19, p. 612). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

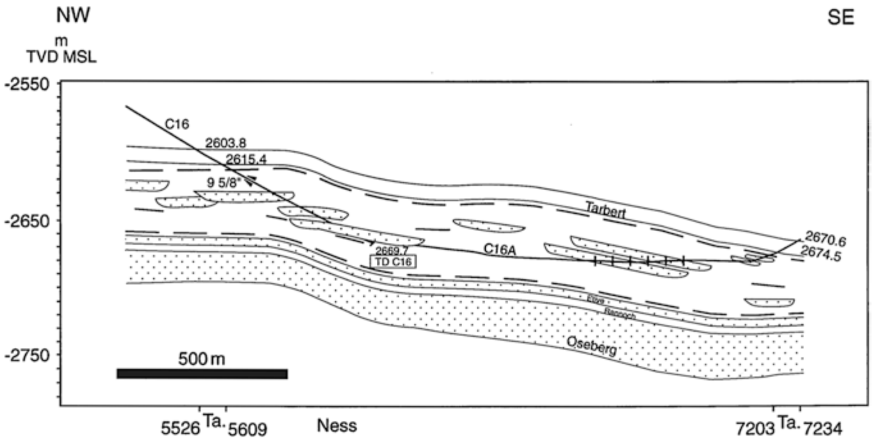


Fig. 3.42 A subhorizontal section through the Ness Formation in the Osberg field, in the Norwegian sector of the North Sea Basin. *Vertical bars* on the well trajectory indicate perforation intervals (Ryseth et al. 1998, Fig. 21, p. 1649). AAPG © 1998. Reprinted by permission of the AAPG whose permission is required for further use

structural features within the study area could potentially contribute to the success of an exploration or production program. Note, however, that there is no discussion here of fluvial style. The geomorphology of the river systems was not considered in these experiments, which clarifies the argument by Larue and Hovadik (2008) that reservoir architecture is of minor importance in assessing reservoir performance. Labourdette's (2011) experiments confirm that *stratigraphic* architecture is important, but have very little to say about fluvial sedimentology. Larue and Hovadik (2008, p. 340) stated:

> For reservoir architecture, two- or three-level designs may not do justice to the entire range of uncertainty including for example, uncertainties in channel width, sinuosity, orientation, sand fairway shape, bar geometry, non-stationarity (that is, trends in sand accumulation), shale heterogeneities and mud-drapes. Moreover, creating geologically realistic reservoir models while honouring well and other conditioning data can be extremely time consuming; it is not uncommon to spend months creating stratigraphically complex 3D facies models.

To refer briefly to just one of the parameters Larue and Hovadik (2008) explored: they examined the effects of varying channel sinuosity on reservoir performance, and demonstrated that where sand bodies had good connectivity, sinuosity was short circuited, essentially eliminating its importance. Additionally, as we noted in Sect. 2.2.4 (Fig. 2.29) the width: depth relationships of preserved sand bodies formed by braided and meandering fluvial systems may be very similar. Larue and Hovadik (2008, p. 368) noted that "details of reservoir architecture are important later in field development, when infill well locations are considered, and residual oil is targeted." We provide some examples of this in Chap. 4.

All of the discussion in this section pertains to the maximizing of production using only traditional vertical wells, for which the geostatistical approach has proved to be the essential tool for developing production models from limited exploration data. The ability of the industry to drill directional wells now creates whole new opportunities. In Chap. 4 we discuss mapping methods, which include such methods as 3-D seismic and surveillance techniques for mapping complex stratigraphic architectures. Figure 3.42 illustrates a 1600-m-long subhorizontal well section through the Ness Formation in the Osberg field, North Sea basin. In a 3-D seismic survey, impedance values were found to provide a reasonably reliable means of discriminating between sand and shale (see Fig. 4.20), and it was on this basis that the course of the well was planned. The cross section shows that the plan was quite successful.

Chapter 4
Basin Mapping Methods

4.1 Introduction

The purpose of this chapter, as for this book as a whole, is not to provide a comprehensive textbook on basin analysis methods as applied to fluvial deposits, but to review recent developments, largely those that have taken place since the publication of "*The geology of fluvial deposits*" in 1996. Formal description, explanation and documentation of methods of facies analysis and architectural analysis and basic subsurface methods (e.g., use of wireline logs) are treated at length in that book, and do not require repetition here. Older, or traditional methods, are discussed in Chaps. 2 and 3 of the present book against the background of our developing understanding of fluvial depositional systems, and how useful these methods are in this evolving context.

Two main themes are examined in this chapter, the mapping of outcrop and shallow subsurface deposits, and the mapping of deeply buried fluvial systems.

Outcrop studies of recent and ancient deposits have been used extensively to provide insights into shallow aquifers and have also served as analogues for the exploration of petroleum reservoir heterogeneities. Aquifer studies represent a more direct application of sedimentological studies, because it is generally the actual aquifer that is under investigation, whereas the reservoirs with which outcrop studies are being compared usually represent completely different geological units. Even where it is the same unit being compared, at the surface and in the subsurface, the comparison is being made over distances of tens to hundreds of metres differences in depth and perhaps several or many kilometres distant, and paleogeographic conditions as represented by the surface outcrops and the buried reservoir could be substantially different. This point was made in Sect. 2.2.1 (see Fig. 2.11) with regard to the problem of defining fluvial style from limited data.

As noted by Geehan (1993, p. 56: see complete quote at the end of Sect. 2.2.2) "outcrops are the only source of geological analogue data that show indisputably what is preserved in the geological record." For this reason, in this book we do not discuss surface or shallow geophysical (e.g., GPR) studies of modern rivers as analogues for

A. Miall, *Fluvial Depositional Systems*, Springer Geology,
DOI: 10.1007/978-3-319-00666-6_4,

fluvial reservoirs. GPR studies may reveal that the surface form of the river is superimposed on older channels and bars, with little modification of their architecture.

Quaternary fluvial sands and gravels commonly constitute important local aquifers, and much work has been carried out to document their porosity–permeability architecture. Increasingly, the two- and three-dimensional data obtainable from outcrop studies is being supplemented by data derived from ground-penetrating radar (GPR), which is a relatively inexpensive technique for generating three-dimensional subsurface remote sensing imagery to depths of up to about ten to twenty metres. Similar work is being done on more ancient deposits to provide analogue data for the interpretation of more deeply buried aquifers and for reservoir analogue studies. For example, outcrop and GPR information may be very useful in the documentation of the types of barriers and baffles to flow generated by mudstones, poorly sorted deposits, abandoned channel-fill clay plugs, etc., of the type illustrated in Fig. 3.35. These mapping procedures are discussed in Sect. 4.2.

The mapping of the deep subsurface is a field of great importance to the reservoir geologist and production engineer (Sect. 4.3). As is argued in Chaps. 2 and 3, mapping methods based on facies model concepts are of limited use at the development phase of petroleum exploitation. The variability in fluvial styles and the lack of definitive criteria for the determination of the scale and architecture of subsurface deposits puts limits on the use of these methods for the planning of primary well placements or enhanced recovery projects. The architectural-element approach may generate useful snapshots of a fluvial system and may be used, as was the facies-model approach in earlier years, to provide analogue information for subsurface development, but the main usefulness of this type of data is to establish architectural and porosity–permeability parameters for the constituent cells in a geostatistical model.

Subsurface exploration and development can now routinely take advantage of directional drilling techniques, and several new methods of subsurface mapping may be used to guide the drill (Table 1.1). Under ideal circumstances, the drill may now be targeted at specific porous subsurface units and may be directed to pass through them horizontally in order to maximize the surface area of the porous unit exposed for fluid flow into the production well. Under these conditions, the usefulness of the geostatistical approach diminishes, while the need for deep sedimentological understanding of the target unit increases. Many aids to the investigation of reservoir compartmentalization have been developed, which can supplement and correct the local sedimentological model. These methods are also described briefly in this chapter, in Sect. 4.3.

4.2 Surface and Shallow Subsurface Architecture

4.2.1 Outcrop Characterization

In their introduction to a special issue of *Sedimentary Geology* on aquifer sedimentology, Huggenberger and Aigner (1999, p. 179) stated:

> Progress towards a better understanding of groundwater circulation and transport processes in aquifers demands a multidisciplinary approach to a host of unresolved problems.

> Although much progress has been made within recent years in interpreting the dynamic character of groundwater systems, many key issues remain to be addressed. In particular, several areas demand attention: the role of sedimentological information (heterogeneity) in groundwater and transport models, the scaling-up of observations from outcrop scale to larger scales and the integration of geological and geophysical information of different quality into the description of an aquifer structure. Still nowadays many of the heterogeneities cannot be recognized directly because of the limitation of measurement techniques.

One of the purposes of their special issue was to explore the role of lithofacies and architectural-element analysis in the examination of aquifer heterogeneity. A good example is the analysis presented in Fig. 4.1. The hydraulic conductivity

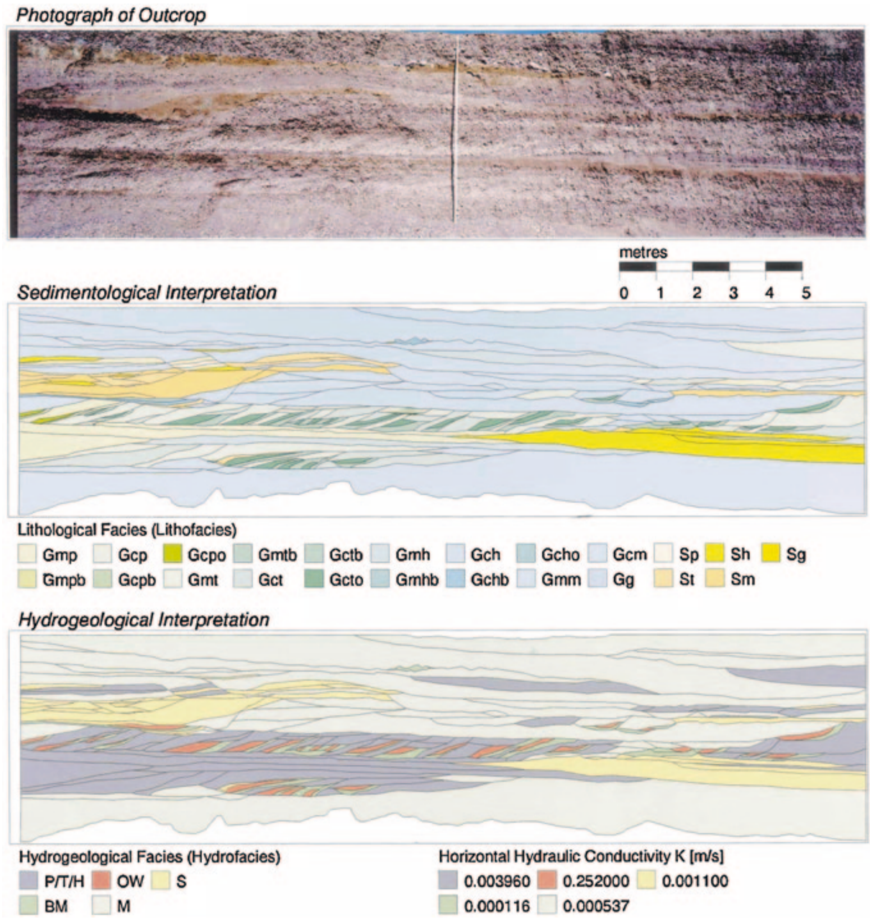


Fig. 4.1 Interpretation of a Quaternary gravel outcrop (*top*) in terms of lithofacies (*centre*) and hydrofacies (*bottom*). Lithofacies classification (*centre panel*) is expanded from Miall (1978) and uses the first letter to indicate grain size (G = gravel, S = sand). Lithofacies may be grouped into five different facies of hydrogeological significance, with uniform hydrogeological parameters: (1) BM: bimodal gravels, (2) OW: open framework gravels, (3) P/T/H: planar/trough/horizontal gravels, (4) M: massive gravels and (5) S: sands (Klingbeil et al. 1999, Fig. 4, p. 307)

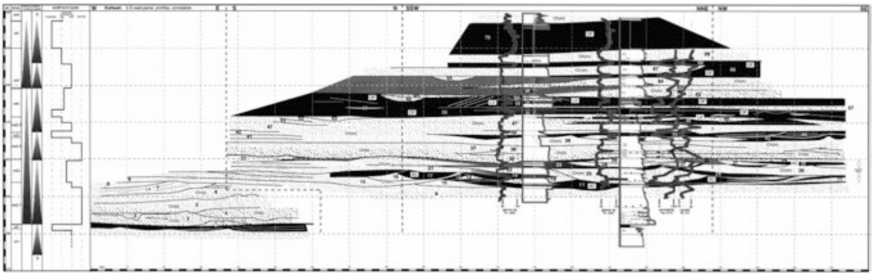


Fig. 4.2 One of a suite of outcrop panels mapped in the Stubensandstein, a Triassic braided stream deposit exposed in southwest Germany. Also shown are the petrophysical logs of two drill holes tied into the outcrop. Lettering on the panel refers to architectural-element codes, listed and explained in Figs. 4.2 and 4.3 (Hornung and Aigner 1999, Fig. 5, p. 221)

values assigned to the hydrofacies in this diagram were derived from "170 in situ and about 50 laboratory gas pneumatic and tracers tests within the single lithofacies units accessible in the outcrops. Furthermore sieve analysis and column experiments with water have been performed." (Klingbeil et al. 1999, p. 304). The purpose of this work was to provide basic ground-truth data for input into geostatistical aquifer models.

In another paper in this special issue, Hornung and Aigner (1999) examined the Stubensandstein, an upper Triassic braided stream deposit interpreted as formed in a terminal alluvial plain setting. In different areas this unit is exploited as a freshwater aquifer, as a site for waste disposal and, downdip, as an oil reservoir. Figure 4.2 illustrates one of their outcrop panels interpreted in terms of local bounding surfaces and architectural elements. Dimensional information for the architectural elements (Fig. 4.3) was derived from numerous outcrop measurements, and porosity and permeability characteristics (Fig. 4.4) were assembled from several hundred laboratory measurements made on core plugs extracted from the outcrops. Gamma-ray profiles were derived by moving a hand-held scintillometer over the outcrops.

Heinz et al. (2003) examined the lithofacies of Quaternary gravel pits in southwestern Germany with a view to determining their hydraulic characteristics. Conductivity values were determined for each facies by experiments in a permeameter column and by empirical calculation. Three types of lithofacies assemblage were identified, as illustrated in Fig. 4.5 and as summarized here (p. 12):

(1) gravel bodies of the "main discharge area", which are dominated by a stacking of large cut-and-fill elements (scour pool fills),
(2) gravel bodies of the "intermediate discharge area", which are characterized by a dominance of accretionary elements (e.g. gravel sheets) and locally small cut-and-fill elements and
(3) gravel bodies of the "minor discharge area", which show an interfingering of many small-scaled accretionary elements with no distinct surface boundaries.

symbol	element	characterization	geometry	lithology sed. str. grain size c s fs ms cs
CH(b)	Channel (bed load)	Coarse-grained sandy bedforms, multilateral and multistory amalgamated channel complexes. Weakly developed fining-up trends.	$n \times 10^{3-4}$ m $n \times 10^{2-3}$ m 1 -10 m	
CH(m)	Channel (mixed load)	Often massive sandbodies. Also alternating layers of silty, fine-grained and coarse-grained sandy bedforms. Clear fining-up trends.	$n \times 10^{2-3}$ m 1-4 m $n \times 10^{1-2}$ m	
CH(s)	Channel (suspended load)	Consists mainly of silt and clay. Rarely thin fine-grained sandy bedforms. No visible fining-up trends.	$n \times 10^{2}$ m 1-9 m $n \times 10^{1-2}$ m	
LA	Lateral accretion	Inclined, alternating layers of silt and clay with fine-grained and coarse-grained sandy bedforms, often irregular bedding contacts. Clear overall fining-up trend.	$n \times 10^{2}$ m 1-9 m $n \times 10^{1-2}$ m	
AC	abandoned channel	Consists mainly of silt and clay. Rarely thin fine-grained sandy bedforms. No visible fining-up trends. Could be reactivated as a channel.	3-50 m 1-9 m 3-50 m?	
LV	Levee	Inclined layers of sand, alternating with silty fine sands. Often overall coarsening-up trend.	$n \times 10^{1-2}$ m 0,5-3 m 5-40 m	
CS	Crevasse splays + sheet floods (LS)	Very coarse to fine sands. Could be amalgamated to thicker packages. Ripple crossbedded or low-angle crossbedded. Bedforms mostly missing. Mostly clear fining-up trends.	$10 - n \times 10^{2}$ m 0,05 - 2 m 20 - 100 m	
FF	Floodplain, Paleosols Overbanks	Horizontal laminated clay and silt. Contains ± developed paleosols, desiccation cracks.	$n \times 10^{3-4}$ m $n \times 10^{3-4}$ m 0,2 - 10 m	
LC	Lacustrine sediments	Dolomitic limestones, dark clays/silts and submerged crevasse sandsheets, alternating multistorey and multilateral.	$n \times 10^{2-3}$ m $n \times 10^{2-3}$ m 0,1 - 2m	

Fig. 4.3 Architectural elements in the Stubensandstein (Hornung and Aigner 1999, Fig. 11, p. 258)

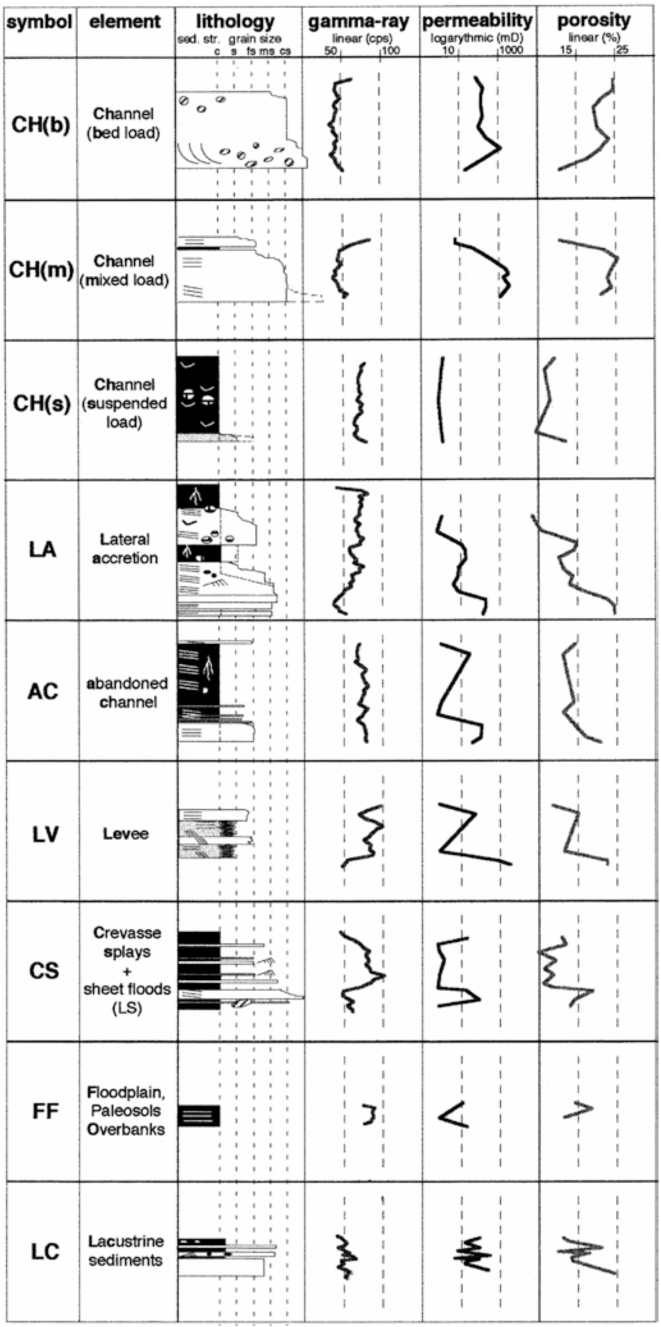


Fig. 4.4 Lithology, petrophysical log character, and porosity–permeability properties of the architectural elements comprising the Stubensandstein (Hornung and Aigner 1999, Fig. 17, p. 269)

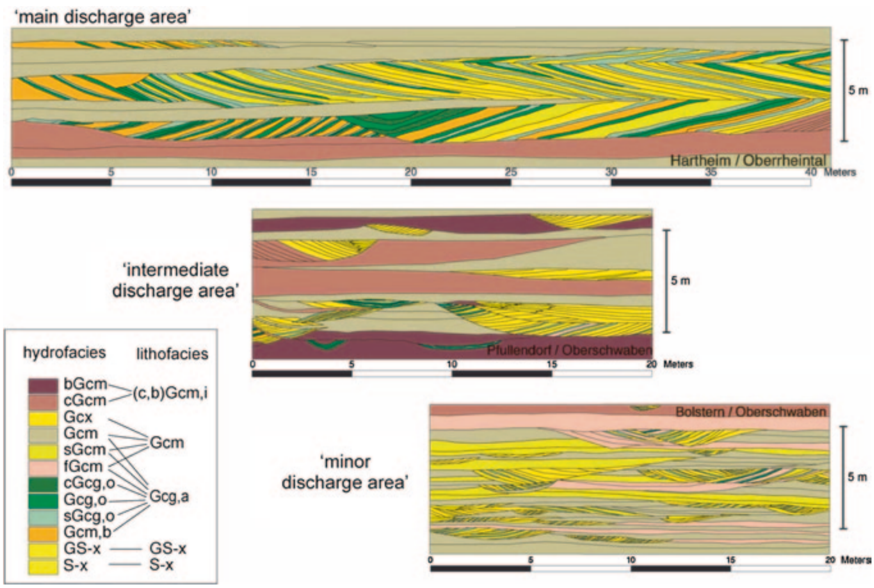


Fig. 4.5 Lithofacies assemblages in Quaternary gravels in SW Germany. The example of the "main discharge area" is oriented parallel to the overall paleoflow direction (from the *left* to the *right*); the example of the "intermediate discharge area" is oriented perpendicular to paleoflow (direction of the outcrop wall) and in the example of the "minor discharge area," paleoflow was from *right* to *left* (Heinz et al. 2003, Fig. 5, p. 13)

The sedimentology of the "main discharge area" is dominated by stacked scour-pool fills, thought to represent the deposits of large, stable channels (Fig. 4.6). The assemblage is a highly complex interfingering and interbedding of permeable lithofacies (cGcg,o/Gcg,o/sGcg,o: essentially clast-supported, graded, open-framework gravels) and less permeable zones (Gcm/sGcm/Gcm,b: clast-supported, massive gravels). These lithofacies have good porosity–permeability characteristics and this assemblage is therefore characterized by high hydraulic conductivity.

The "intermediate discharge area" is the product of small, unstable channels markedly affected by high-magnitude floods, and represented by many cut-and-fill elements.

Lithofacies of the "minor discharge area" are dominated by fine-grained gravels and by poor sorting, suggests flashy discharge, with periods of low flow in minor channels.

Flow modeling of these three types of assemblage indicate that the intermediate and low discharge areas are characterized by the most homogeneous flow patterns. The more highly porous and permeable lithofacies in the high discharge area leads to fluid transmission preferentially through these zones. In extreme cases such units would be termed "thief zones" in petroleum production parlance, because they lead to premature drainage and (in the case of enhanced recovery

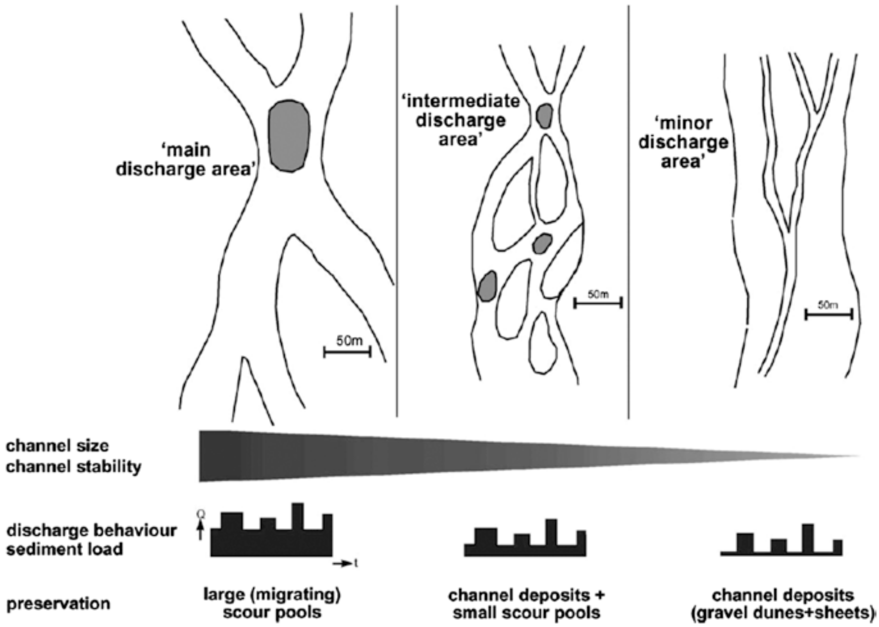


Fig. 4.6 The interpreted fluvial setting of the three main lithofacies assemblages in Quaternary gravels in SW Germany (Heinz et al. 2003, Fig. 7, p. 16)

projects) water breakthrough and the isolation and abandonment of significant oil pockets.

The authors emphasize that their data and interpretations did not adequately take into account the downstream length of lithofacies units. Studies such as theirs, based on limited outcrop data, are commonly limited by these types of considerations.

Comparable outcrop studies have been carried out for the purpose of developing petroleum reservoir models. A good example is that described by Stephen and Dalrymple (2002). They created a photomosaic of a large outcrop approximately 450 m long and 45 m high of the Straight Cliffs Formation (Upper Cretaceous) in Utah (Fig. 4.7). Paleocurrent measurements indicated that the outcrop is nearly normal to paleoflow. The outcrop was straightened in order to eliminate structural complications, and digitized. Units identified as shales or heterolithics—those beds which are assumed to represent what would be impermeable barriers or baffles to flow in the subsurface—were measured and their lengths tabulated. These data were then used as input into flow simulation programs, with porosity and permeability set at various rates to assess reservoir performance.

> The different well configurations show the effect of the heterogeneity distributions on flow. For the case with linear [horizontal] drive, the shale and heterolithic distributions have negligible influence on flow if the sand is considered to be homogeneous and isotropic, once the volumetric variability has been accounted for. Such behaviour is not surprising because the flow is effectively parallel to the alignment of the vertical flow

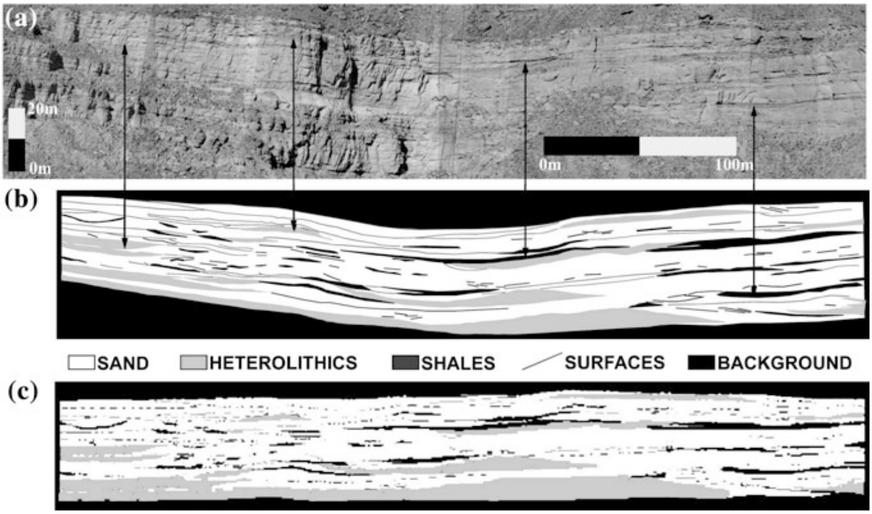


Fig. 4.7 **a** Outcrop of the Straight Cliffs Formation, Utah; **b** Interpretation of the outcrop in terms of basic lithofacies; **c** The outcrop is artificially straightened and digitized using cell sizes of 74 × 61 cm, yielding a total cell count of 661 horizontally and 81 vertically (Stephen and Dalrymple 2002, Fig. 6, p. 803). AAPG © 2002. Reprinted by permission of the AAPG whose permission is required for further use

> barriers and impediments. Low or zero net-to-gross lithofacies do not affect the flow paths or production efficiency and need only be considered in a very small number of geostatistical realizations. If sand permeability is highly variable between sand bodies, however, then the vertical flow properties of shales and heterolithics can have influence but still may be ignored. These lithologies influence the crossflow between high- and low-permeability sand bodies, which slightly affects the sweep efficiency in our case. We found that, for vertical flow due to horizontal wells, the impact of heterolithic and shale flow properties is very strong, assuming homogeneous sand. The flow is strongly influenced by the combination of different types of lithology, which combine to affect the tortuousity of the flow paths. When modeling a horizontal well parallel to the cross section, heterolithics between wide shale pods can improve the sweep efficiency (Stephen and Dalrymple 2002, p. 817).

Figures 4.8, 4.9, 4.10 illustrate a detailed study of the bounding surfaces and architectural elements of the reservoir unit in the Karamay oil field, China, based on studies of nearby outcrops, and the relationship of these sedimentological details to the porosity and permeability patterns displayed by this producing unit. The annotation of the outcrop lithofacies, architectural elements and bounding surfaces uses methods adapted from Miall (1996). Figure 4.8 is the architectural classification developed by Jiao et al. (2005) for the porous, producing sand and gravel units. Outcrop examples of two of the sand complexes are illustrated in Fig. 4.9.

The reservoir heterogeneity has been conceptualized on three scales (Fig. 4.10) comparable to those illustrated in Fig. 2.9. At the microscale range of cores, pore

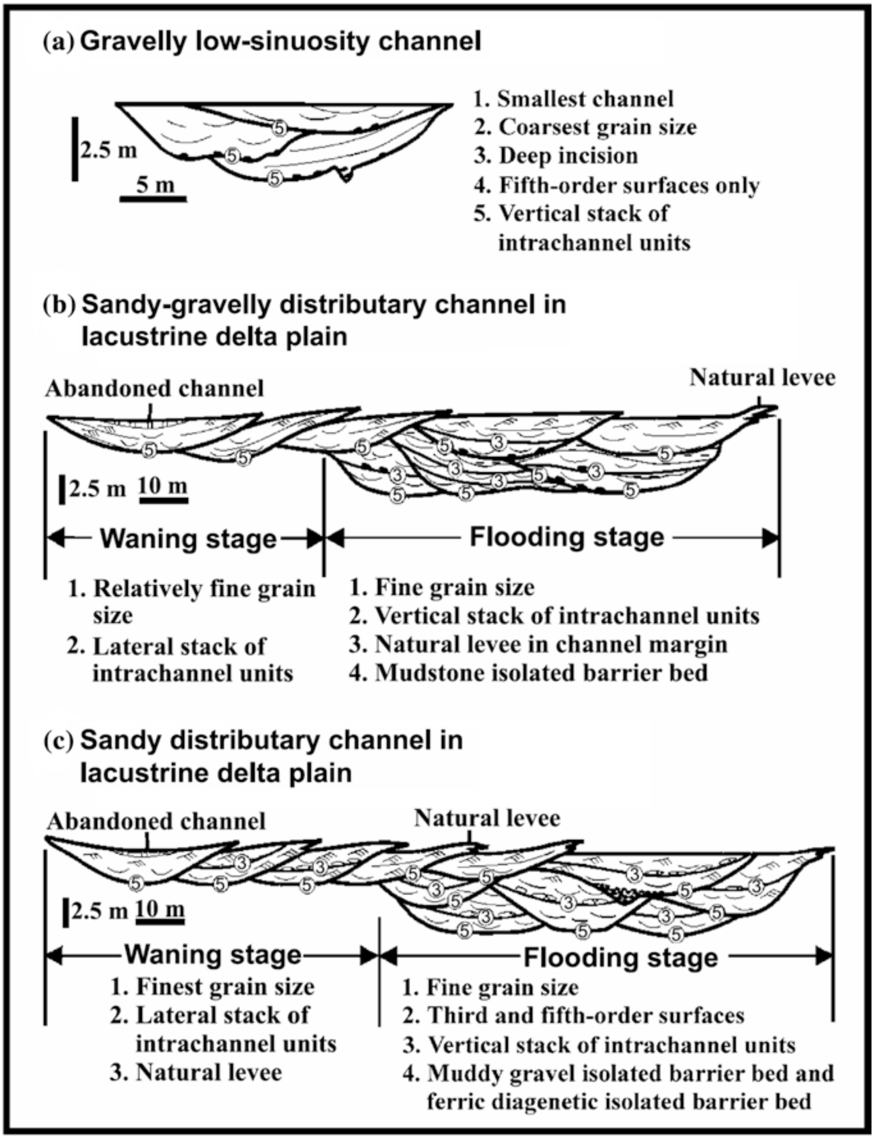


Fig. 4.8 Architectural elements comprising the reservoir of the Karamay oil field, China (Jiao et al. 2005, Fig. 10, p. 537). AAPG © 2005. Reprinted by permission of the AAPG whose permission is required for further use

texture and permeability anisotropy were measured. The greatest horizontal permeability (47.3 md) was found to be parallel to the paleocurrent direction; intermediate values of horizontal permeability (16.4 md) were perpendicular to the paleocurrent direction, and the vertical permeability was found to be the lowest.

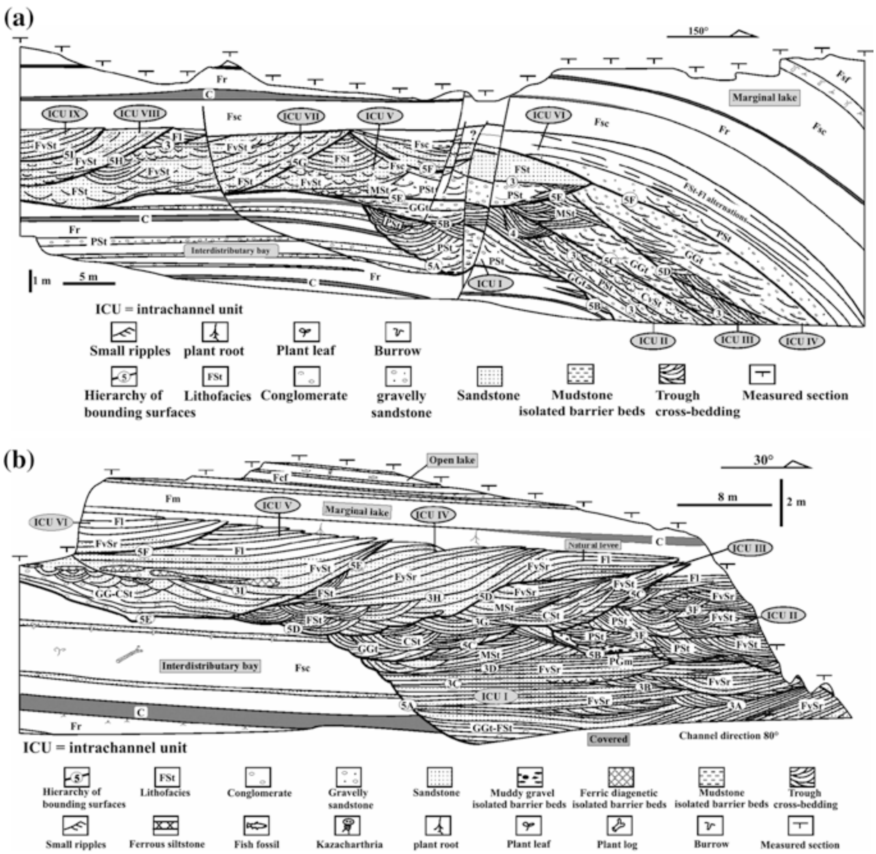


Fig. 4.9 **a** Architectural unit diagram of a sandy-gravelly distributary channel in the expanding-lacustrine systems tract of the Karamay Formation; **b** Architectural unit framework of a sandy distributary channel (Jiao et al. 2005, Fig. 7, p. 535 and Fig. 9, p. 536). AAPG © 2005. Reprinted by permission of the AAPG whose permission is required for further use

In the middle-scale heterogeneity range, that of individual channels, isolated fluid barrier beds were identified at different bounding surfaces (Fig. 4.11). Porosity and permeability are also highly dependent on lithofacies. Jiao et al. (2005, p. 539) identified three types of fluid barrier that could be used to predict the distribution pattern of the fluid-flow units and their internal porosity and permeability, and to make a genetic interpretation of all these phenomena (figure numbers cited below have been changed to those used in this book):

> (1) Fine-grained isolated barrier beds, sediments of intrachannel units or of macroform accretion units resulting from energy attenuation at the end of deposition, were distributed mainly at the tops of these sedimentary bodies. This kind of isolated barrier bed was thickened in the direction of the interdistributary bays or even connected with fine-grained sediments in the interdistributary bay. In specific cases, well-connected abandoned channel mudstones may also serve as isolated barrier beds whose geometric shapes were not only

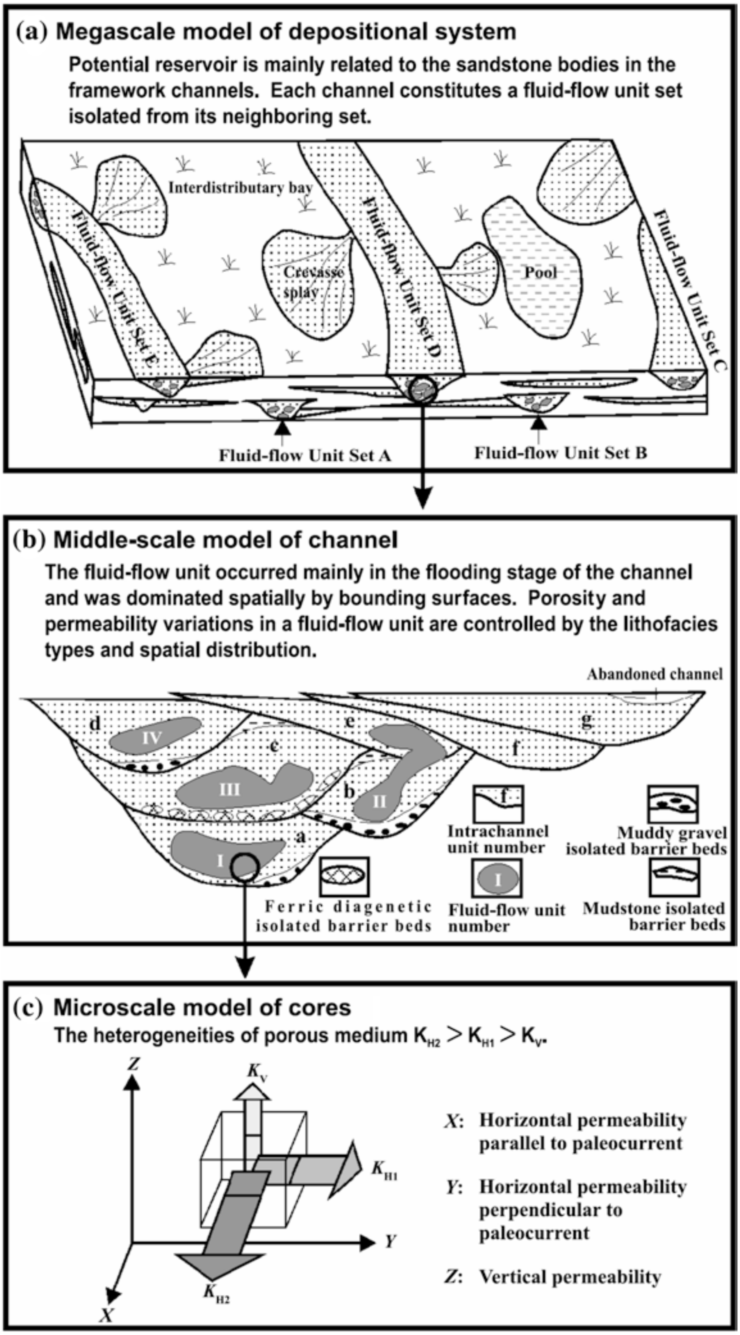


Fig. 4.10 The scales of heterogeneity in the Karamay oil field reservoir (Jiao et al. 2005, Fig. 11, p. 538). AAPG © 2005. Reprinted by permission of the AAPG whose permission is required for further use

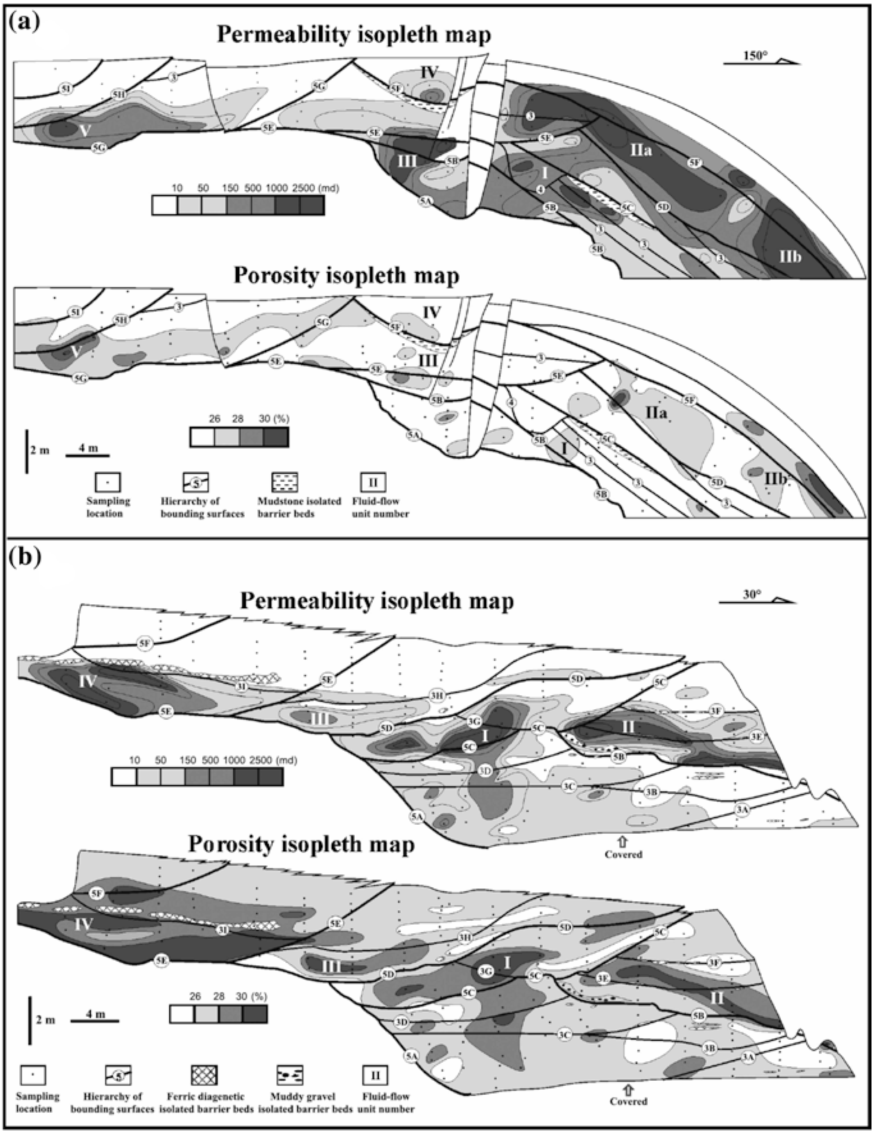


Fig. 4.11 Distribution pattern of porosity and permeability in (**a**) a Sandy-gravelly distributary channel (Fig. 4.9a), and (**b**) a sandy distributary channel (Fig. 4.9b), in a lacustrine delta-plain distributary system, Karamay oil field (Jiao et al. 2005, Fig. 13, p. 541). AAPG © 2005. Reprinted by permission of the AAPG whose permission is required for further use

controlled by deposition but also closely associated with the later scouring (Fig 4.11a). (2) Muddy gravel isolated barrier beds, actually a kind of relatively concentrated channel lag, overlay fifth-order or third-order surfaces and thin out toward both sides of the channel (Fig 4.11a). (3) A ferric diagenetic isolated barrier bed was a heterogeneous layer of nodules along a sedimentary bounding surface formed during the diagenetic process (Fig 4.11b). The porosity and permeability of the three isolated barrier beds mentioned above were extremely poor (e.g., 2.66 %, 24.9 md for a muddy isolated barrier bed, 1.47 %, 26.0 md for a muddy-gravelly isolated barrier bed, and 6.49 %, 27.2 md for a ferric diagenetic isolated barrier bed).

In the megascale range of the depositional system, porosity and permeability were compared between different genetic facies to identify high quality reservoirs.

4.2.2 Ground-Penetrating Radar (GPR)

Many of the architectural studies reported on in previous sections of this book (and in Miall (1996)) have been essentially two-dimensional in character; that is, they are based on studies of outcrops that may be high, stratigraphically (metres to tens of metres) and lengthy (tens to hundreds of metres) but which display little or no depth to provide architectural information in the third dimension. In a demonstration study of some ancient fluvial bar forms, Miall (1994) demonstrated how the measurement of surface orientations at outcrop, including channel margins, accretion surfaces and crossbedding, could provide useful insights into the third dimension immediately behind the outcrop surface, but such information is necessarily limited in its applicability. This has been repeatedly pointed out (e.g., Bridge 1985, 1993, 2003) as a criticism of outcrop methods.

The development of ground-penetrating radar methods seemed to point to a way around this problem. This is a geophysical method that uses radar pulses to image the subsurface, using electromagnetic energy in the microwave band. Superficially, the technique is comparable to reflection-seismic methods, in that reflected energy is collected, processed and displayed, typically in the form of two-dimensional cross-sections, where the *x*-axis corresponds to the track of the radar survey and the *y*-axis represents two-way travel time. The technique was first developed in 1929 to study the depth of a glacier (http://www.g-p-r.com/introduc.htm) but applications to geological studies did not commence until the 1970s.

GPR offers a potential resolution in the centimetre range for subsurface studies, and the ideal combination for architectural studies is where large two-dimensional outcrops occur within horizontal to gently dipping strata, where the top surface of the outcrop is developed within one of the layers of the succession, affording a flat surface over which to run the survey (such as the mesas of the American southwest). A test and calibration line may be run a short distance behind the outcrop, to provide a basis for direct outcrop-to-radar interpretation (Fig. 4.12), and a grid of intersecting lines can then be run to extend the interpretation back into the subsurface and to provide extensive information on the third dimension. The utility of the technique is limited primarily by the degree of water saturation of the surveyed rocks, because the radar signal is rapidly attenuated in water-saturated media. The signal is also absorbed by clay layers. Useful reflections are typically not returned from below the water table, and so as far as surveying ancient rocks is concerned, GPR is most useful in dry environments, such as the American southwest, where penetration may exceed 10 m.

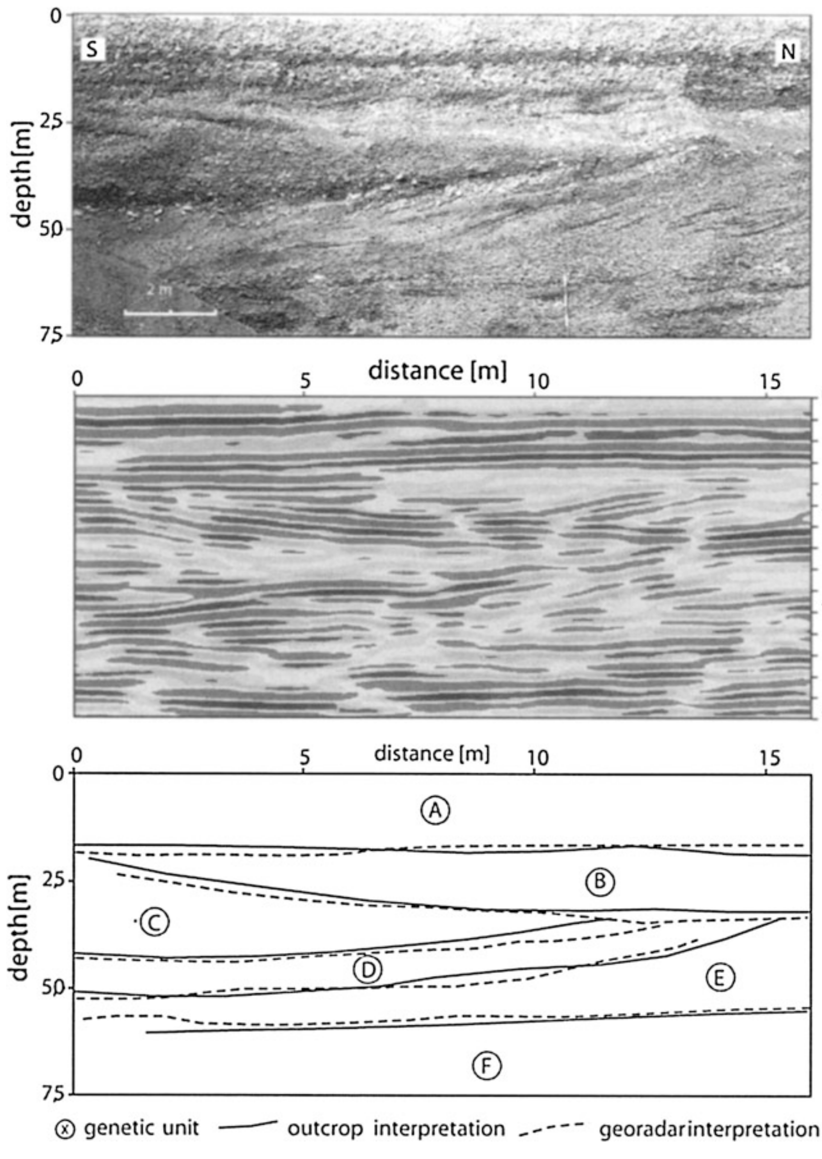


Fig. 4.12 Comparison of outcrop-wall photograph and radar image of a Quaternary fluvial gravel, southwestern Germany (Heinz and Aigner 2003, Fig. 3, p. 103)

A compilation of GPR studies was edited by Bristow (2003), and includes a practical guide to field methods (Jol and Bristow 2003). Several studies of fluvial deposits are included in this collection, and are referenced in this section. A major review of GPR and its uses was provided by Neal (2004). This study focuses primarily on geophysical principles and methodology, and while the author makes reference to sequence and facies studies, the only practical examples described are of modern coastal strandplains and sand dunes. Akinpelu (2010) carried out a wide-ranging review of GPR studies and carried out a suite of field tests, which lead to a number of important conclusions about the methodology, discussed below.

This is not the place for a detailed description of field methods and techniques. However, some essential points need to be made for the purpose of this review:

> Most sedimentary studies use antennae with frequencies between 50 and 500 megahertz (MHz), with the majority of research reported in this volume using 100 MHz antennae. Higher frequency antennae (400 to 1000 MHz) have shorter wavelengths which can yield high resolution but little depth of penetration. Lower frequency antennae (10 to 50 MHz) have longer wavelengths that can yield greater depth of penetration but lower resolution (Jol and Bristow 2003, p. 10).

Akinpelu (2010) reviewed the range of techniques used in GPR studies of clastic sedimentary rocks, focusing on fluvial deposits. One of the most important attributes of the equipment is the frequency of the antennae used to transmit and receive radar energy.

> As observed in many of the outcrop locations studied in this research; significant signal attenuation and low penetration depth are the primary limitations of GPR imaging technology to outcrop studies. There is virtually nothing that can be done to reduce signal attenuation due to shale content and the presence of diagenetic clay in the formation being studied or overlying formations. Deeper signal penetration can be achieved by using lower frequency antennae (12.5–100 MHz) for GPR surveys although lower frequency antennas are usually bulkier …. which implies longer data acquisition time and in some cases, additional survey crew. Resolution will also be sacrificed with lower frequency antennas but the lower higher end of the high frequency antennas (100–400 MHz) is still ideal for resolving architectural elements and macroforms. (Akinpelu 2010, p. 151)

Akinpelu (2010) noted that an antenna frequency of 12.5 MHz is limited to less than 50 m penetration depth in most clastic rocks, even in dry clay-free lithologies. A 100 MHz antenna commonly is used for deep (10–20 m) penetration.

Step-length—the spacing between individual ground readings—is of critical importance for the imaging of three-dimensional structures in the subsurface, such as channels, accretion surfaces and crossbedding (Figs. 4.13, 4.14). Jol and Bristow (2003, p. 12) noted that a step length of 1 m is commonly used with an antenna frequency of 100 MHz. However, structures less than about 2.5 m wide will not be imaged completely at this spacing.

Useful three-dimensional information may be obtained from a grid of lines run at right-angles to each other (Figs. 4.15, 4.16). However, the full power of three dimensional analysis requires considerably more data. Line spacing will range from 10 cm for a 250 MHz antenna, for very high resolution studies, to 5 m with a 12.5 MHz antenna, for greater penetration but lower resolution (Akinpelu 2010, p. 150). An example is shown in Fig. 4.17.

In geological applications, the parameters which have the greatest influence on the electrical properties, and hence on penetration depth, velocity and reflector characteristics are: (i) water saturation, (ii) clay content and (iii) pore water salinity. Dry, clay-free sediments generally have higher velocities of propagation

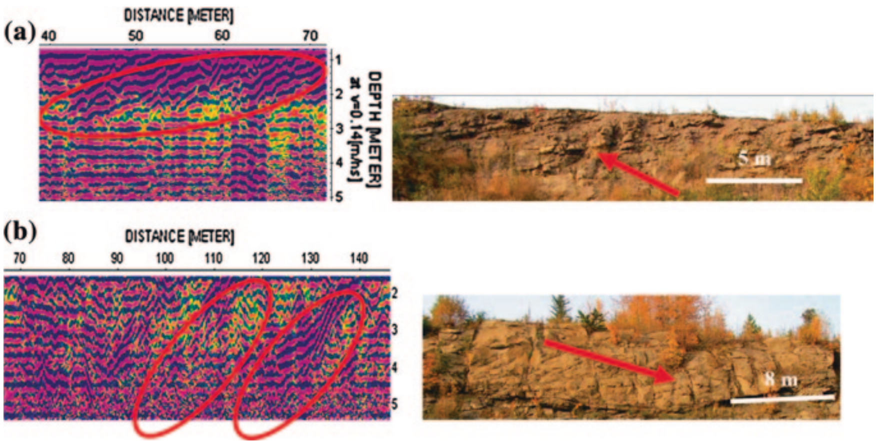


Fig. 4.13 Macroscopic features of an ancient fluvial deposit, as seen in outcrop and as revealed by a GPR survey behind the outcrop. **a** Inclined reflectors dipping at 10–20°, resting on flat reflector, interpreted as inclined heterolithic stratification (IHS: Thomas et al. 1987); **b** Moderately dipping (20–30°) reflectors showing concave-upward downlap onto basal surface, interpreted as lateral accretion deposits within a channel. Dunvegan Formation (Upper Cretaceous), Pink Mountain, British Columbia (Akinpelu 2010)

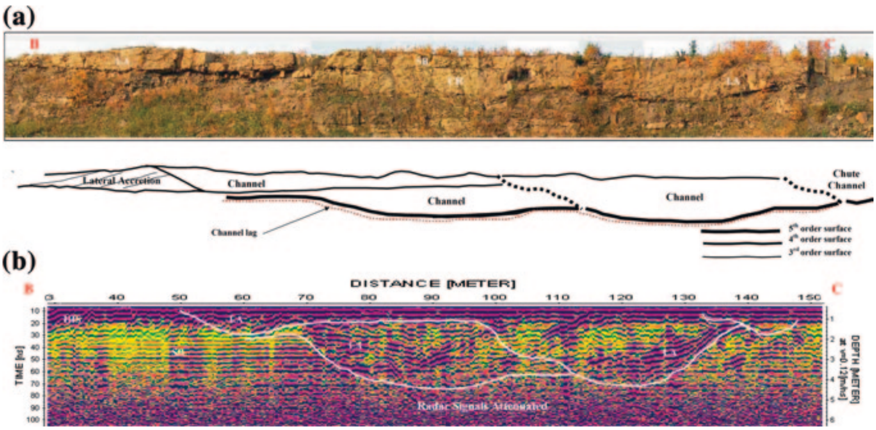


Fig. 4.14 A suite of intersecting channels. Dunvegan Formation (Upper Cretaceous), Pink Mountain, British Columbia (Akinpelu 2010)

and lower attenuation coefficients than wet strata. Depth of penetration is therefore highest in dry, porous sediments (e.g. limestone, where probing depths may be up to 100 m) and lowest in water-saturated clay (where depth of penetration may be <1 m).

Definition of radar facies is often based on shape of reflections; dip of reflections; relationship between reflections, reflection continuity and reflection amplitude (Neal 2004);

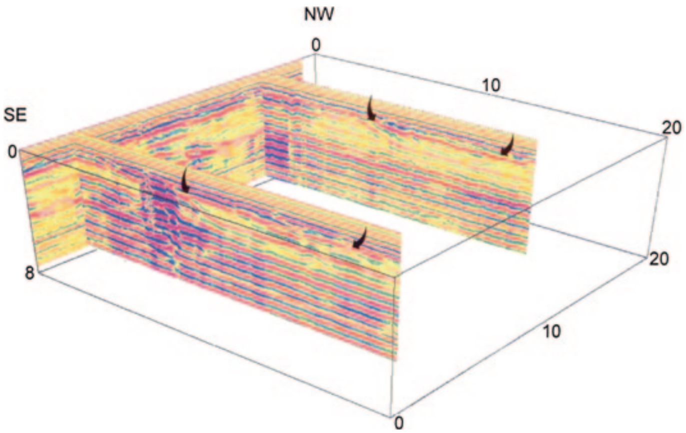


Fig. 4.15 A grid of GPR lines through an interpreted crevasse-splay sheet, eroded into flat-bedded floodplain deposits. Keuper Sandstone (Upper Triassic), near Tübingen, southern Germany (Aigner et al. 1996, Fig. 6b, p. 402)

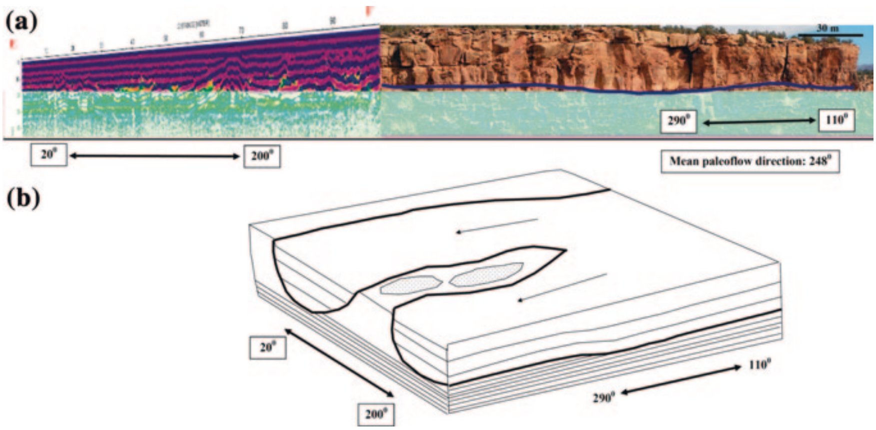


Fig. 4.16 **a** At *right*: an outcrop of the Shinarump Member of the Chinle Formation (Triassic), Hurricane Mesa, Utah. A GPR survey line was run at right-angles to the outcrop, as shown at *left*. **b** Block diagram, based on outcrop and radar line, showing interpretation of channel with mid-channel bar form (Akinpelu 2010)

these criteria can be significantly influenced by the processing steps. Continuous reflection can appear discontinuous by applying inappropriate gains during data processing and reflection dips can result from spurious signals; hence the importance of proper signal processing steps and outcrop data as ground truth to verify the fidelity of radar data. This also underscores the importance of documenting data acquisition and processing steps in GPR publications to allow comparison between studies. (Akinpelu 2010, p. 33)

Because GPR data interpretation depends significantly on the data processing steps employed; more involvement of geophysicists on projects involving three dimensional outcrop studies is required especially as GPR data processing, GPR attribute analysis and visualization gets increasingly advanced. (Akinpelu 2010, p. 148)

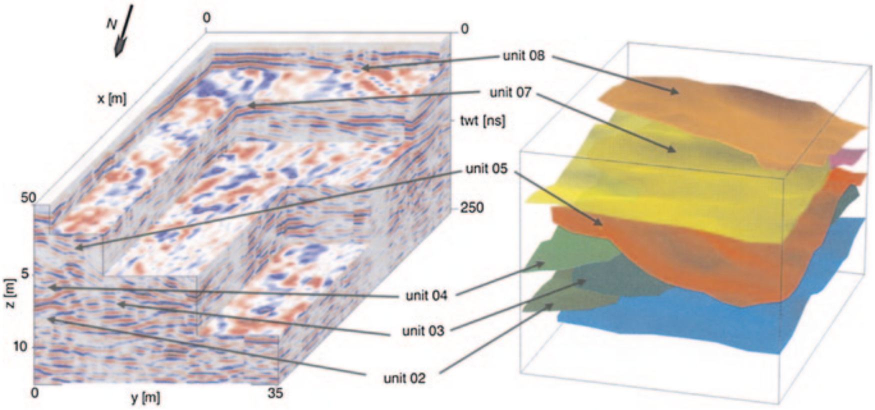


Fig. 4.17 A three-dimensional GPR survey through a Quaternary fluvial gravel, southwestern Germany, showing an interpretation in terms of three-dimensional macroforms (Heinz and Aigner 2003, Fig. 5, p. 105)

Several authors have proposed systems of "radar facies" whereby particular reflection configurations are said to characterize certain depositional fabrics. However, Akinpelu (2010, pp. 66–68) warned against an over-simplified approach to this topic, arguing that reflection characteristics depend on antenna frequency, step length, ground conditions and data processing procedures.

Amongst applications to fluvial facies are the following: Studies of Quaternary gravels: Huggenberger (1993), Asprion and Aigner (1999), Beres et al. (1999), Heinz and Aigner (2003), Dubreuil-Boisclair et al. (2011); shallow subsurface mapping of a modern river and its deposits: Fielding et al. (1999), Lesmes et al. (2002), Lunt and Bridge (2004), Lunt et al. (2004), Sambrook Smith et al. (2006). There have been a few studies of ancient fluvial deposits: Gawthorpe et al. (1993), Aigner et al. (1996), Stephens (1994) and of fluvial distributary channels: McMechan et al. (1997), Szerbiak et al. (2001), Hammon et al. (2002), Corebanu et al. (2002), Zeng et al. (2004). Other GPR studies of modern and ancient clastic deposits are listed by Akinpelu (2010).

With appropriate antenna frequency and step length, significant details of macroscopic fluvial architecture may be revealed by GPR studies. Akinpelu (2010) carried out a series of case studies of fluvial units to explore this capacity of GPR, and a few examples are illustrated in Figs. 4.13 and 4.14. Grid-line surveys may reveal the subsurface extent of major tabular, sheet-like, channel-shape or other bodies (Figs. 4.15, 4.16, 4.17). This may provide important information on the subsurface orientation and scale of channels and other complex macroform units (Stephens 1994).

Many studies have been carried out on clastic units with the object of developing three-dimensional information on porosity–permeability architecture (McMechan et al. 1997; Szerbiak et al. 2001; Hammon et al. 2002; Corebanu et al.

2002; Zeng et al. 2004). In these studies, actual facies characteristics are of less importance than the scale and distribution in three dimensions of potential reservoir volumes, for input into flow modeling exercises.

4.3 Mapping the Deep Subsurface

4.3.1 Seismic Methods

The routine use of three-dimensional seismic methods in exploration and development has revolutionized the application of sedimentology to the subsurface mapping of complex stratigraphies. The methods were pioneered by Alistair Brown and published in AAPG Memoir 42, now in its seventh edition (Brown 2011). Developments in the sedimentological applications of 3-D seismic were led by Henry Posamentier, and have recently been brought together in a very useful compilation of regional studies (Davies et al. 2007). An introductory article by Posamentier et al. (2007) explains the basis of what has come to be called *seismic geomorphology*—the interpretation of landforms and depositional systems from seismic data. Horizontal seismic sections (also called *seiscrops*) are essential for this purpose. Three-dimensional methods provide the ability to construct horizontal sections through the data volume at any selected level, and such sections may also be drawn along stratigraphic surface defined by the user (termed *stratal slices*), which thereby provides the ability to account for structural dip, drape, etc., and adjust a stratigraphic surface back to its original horizontal attitude. This also permits the user to reduce complications in the reflection record from gently dipping, closely-spaced horizons. Such images of the data are likely to be much more useful than conventional vertical sections because they display depositional elements in a landscape form that is familiar to anyone who has spent some time looking at modern depositional processes. The issue of seismic resolution is also a potential problem with the examination of vertical sections. Resolution (the ability to resolve thin beds—conventionally regarded as 1/4 of the wavelength of the seismic wave) ranges between about 10 m under ideal conditions at shallow exploration depths, to more than 200 m at depths of several kilometres. Integration of seismic data with borehole information, including wireline log characteristics and, where available, sample and core data is essential for a full description and interpretation of the depositional system. Wireline logs can add useful information on lithologic variability especially where vertical seismic resolution is poor, and provide quantitative data for other, related calculations, as noted in the following discussion.

Two figures illustrate the power of the technique. Figure 4.18 illustrates a meandering channel system developed along a depositional surface that has undergone gentle folding (this could simply be a drape effect reflecting differential compaction). The vertical section (in grey tones) illustrates the structure; the stratal slice (in colour) is a map of amplitude variations, the darker colours (reds and blues) highlighting what is in this case a meandering turbidite channel in the Gulf

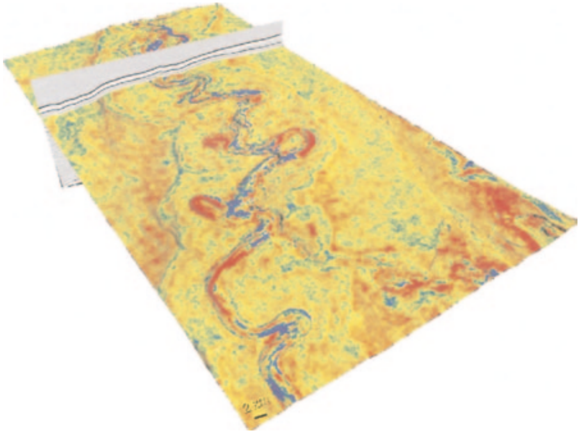


Fig. 4.18 A map of amplitude variations draped on a gently dipping stratal surface, illustrating a meandering turbidite channel in shallow sediments in the Gulf of Mexico (Posamentier et al. 2007, Fig. 2, p. 3)

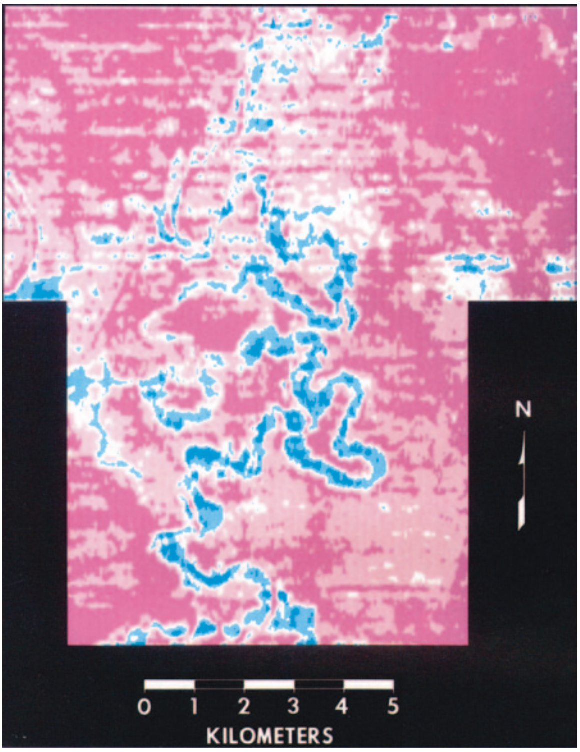


Fig. 4.19 A meandering fluvial channel in shallow sediments in the Gulf of Thailand. From the first (1986) edition of Brown's AAPG memoir (Brown 2011 is the latest edition). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

of Mexico. Note the use of a perspective view for illustrating this depositional system, simulating in some ways what a landform would look like as viewed out of a low-flying aircraft window. A meandering fluvial system is illustrated in Fig. 4.19. This was one of the first illustrations of a depositional system to be imaged in this way; it appeared in the first (1986) edition of Brown's memoir.

Ethridge and Schumm (2007) provided an overview of fluvial channel styles and classification, and made reference to some early examples of seismic-geomorphic interpretations.

The value of a detailed and precise architectural reconstruction of the subsurface may be illustrated by Fig. 4.20, a diagram from an exploration program completed now more than a decade ago (techniques have developed significantly since that time). The section clearly indicates where productive reservoir intervals are expected to be located. A comparison with Fig. 3.42 indicates how successful the prediction was.

Sarzalejo and Hart (2006) illustrated several examples of meandering systems in the Mannville Group (Lower Cretaceous) of southeastern Saskatchewan (Figs. 4.21, 4.22, 4.23, 4.24). Seismic resolution (1/4 wavelength) is approximately 12 m in these examples. Many gamma-ray logs are available from positions within the selected seismic volumes, but no core or sample data are obtainable. The combination of horizontal and vertical seismic sections and gamma-ray logs provides a powerful tool for the interpretation of depositional systems which, in the appropriate settings, may enable exploration to focus directly on prospective units.

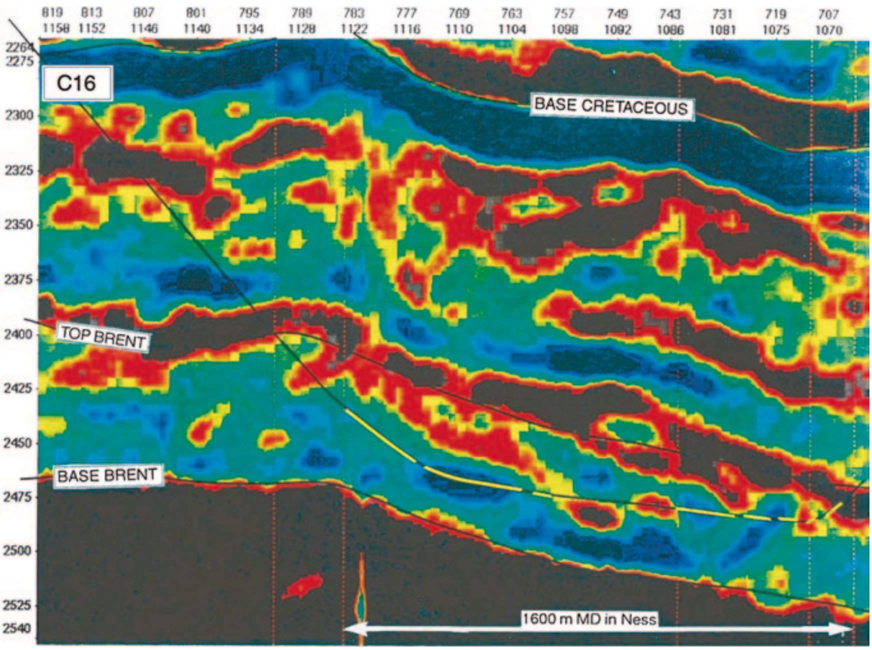


Fig. 4.20 Three-dimensional seismic section along the proposed path of well 30/6-C16 (see Fig. 3.37). Low impedance values (*green–blue* colors) along the well trajectory indicate reservoir sandstones (*yellow* parts of the trajectory). The impedance variation indicates that the well would penetrate two main sandstone-dominated sections separated laterally by a mudrock-dominated interval (Ryseth et al. 1998, Fig. 20, p. 1648). AAPG © 1998. Reprinted by permission of the AAPG whose permission is required for further use

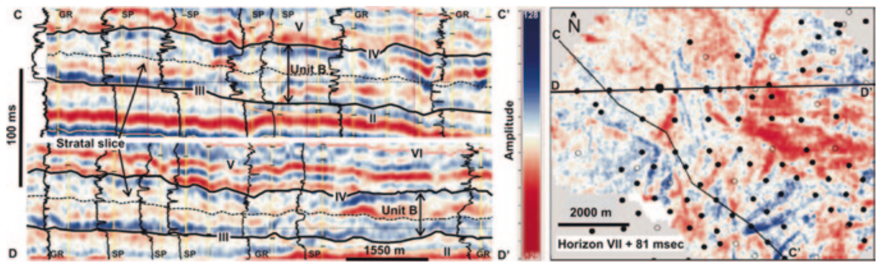


Fig. 4.21 Meandering fluvial system, Manville Group, southeastern Saskatchewan. From Sarzalejo and Hart (2006, Fig. 10, p. 147)

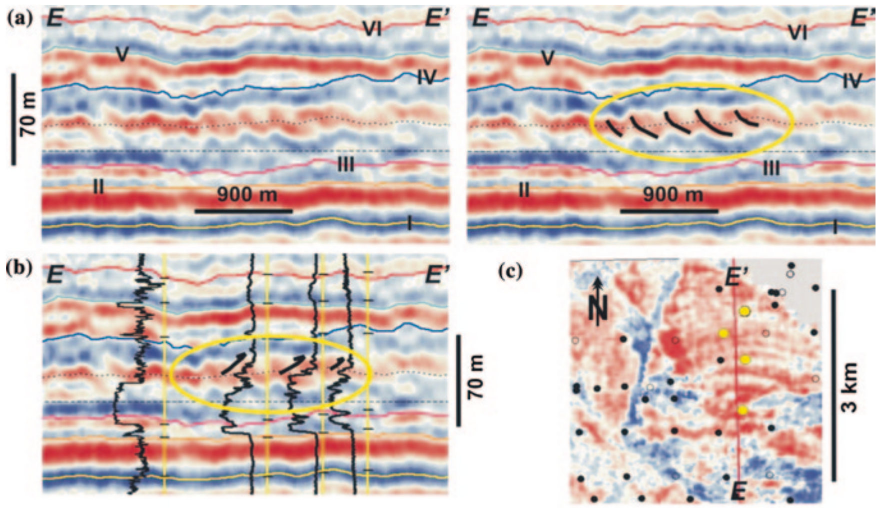


Fig. 4.22 Meandering fluvial system, Manville Group, southeastern Saskatchewan. From Sarzalejo and Hart (2006, Fig. 11, p. 147)

The time-slice map in Fig. 4.21 shows bands of darker colours (reds and blues) interpreted as sets of scroll bars of intersecting point bar complexes. The vertical section, with the gamma-ray logs, illustrate the classic serrated bell-shaped signature of upward-fining fluvial units, ranging between about 50 and 70 m in thickness.

Figure 4.22 provides an even clearer example of a meandering system. The vertical section includes a suite of shingled reflectors between numbered surfaces III and IV, that are highlighted by a yellow ellipse in the interpreted section (at top right). These are interpreted as lateral-accretion surfaces within a point bar. Their appearance in such a section indicates a certain degree of lithologic heterogeneity in the point bar, such as sandstone intervals interbedded with significant mud drapes or surfaces of internal erosion (3rd-order surfaces) marked by strongly contrasting lithology, such as poorly sorted, muddy or pebbly deposits. These surfaces

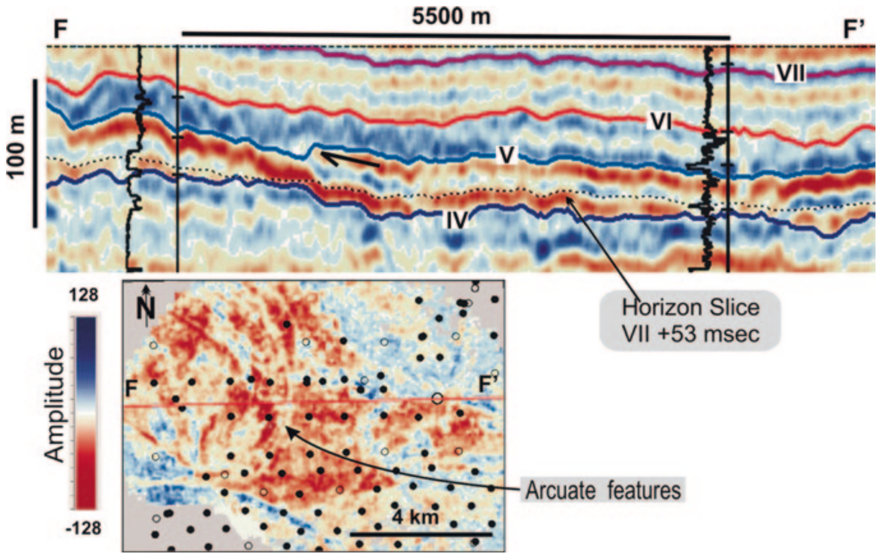


Fig. 4.23 Meandering fluvial system, Manville Group, southeastern Saskatchewan. From Sarzalejo and Hart (2006, Fig. 12, p. 148)

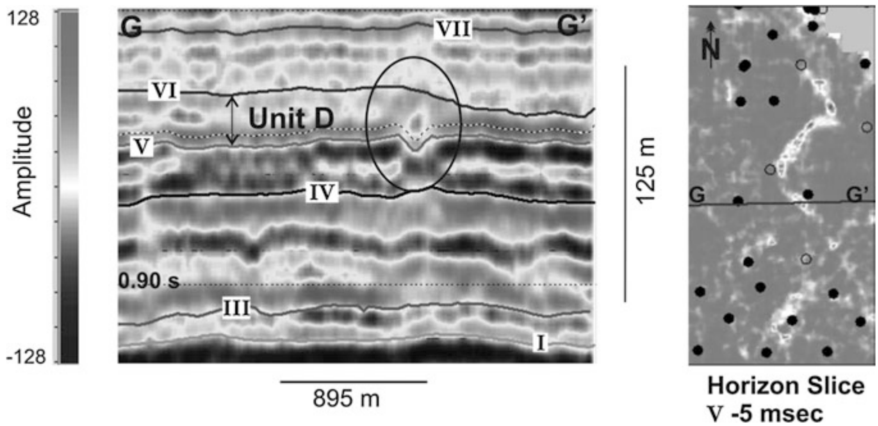


Fig. 4.24 A single-thread fluvial channel, Manville Group, southeastern Saskatchewan. From Sarzalejo and Hart (2006, Fig. 14, p. 150)

also show up as serrations in the gamma-ray logs, and the time slice (Fig. 4.22c) clearly shows curved scroll-bar traces, the direction of curvature indicating point-bar migration and growth toward the north.

Figure 4.24 illustrates an example of a single-thread channel about 140 m wide. The curved traces underlying this channel probably reflect compaction of fine-grained floodplain deposits beneath the denser and less compactible sand fill of the channel.

The seismic sections illustrated in Figs. 4.21, 4.22, 4.23, 4.24 reveal an unusually high resolution level. Two additional figures (Figs. 4.25, 4.26) provide cross-section at an even higher resolution. These are from a sparker-seismic survey of the shallow section immediately beneath the sea floor in the Gulf of Thailand (Reijenstein et al. 2011). The identification of these sections as point bars is established by 3-D seiscrop sections of the same rock volume. Sparker surveys are only possible in shallow sections but provide a vertical resolution approximately 25 times that of conventional 3-D petroleum seismic, at about 25 cm. Figure 4.25 illustrates a typical point-bar cross-section, complete with chute channels and an abandoned-channel mud plug. Figure 4.26 illustrates the bidirectional dip that develops around the "point" of a point bar, with accretionary dip oriented toward the channel all around the meander bend.

Skilled interpreters can extract even more information about depositional processes. Figure 4.27 illustrates a vertical seismic section in which laterally-limited channel sands have been identified as low-amplitude (red) patches in the section. Three stratal slices taken within a 12-ms window show the subtle shifts in channel position with time (Fig. 4.28). These can be interpreted in terms of the evolution of a meander belt (Fig. 4.29). As expected, the meanders show evidence of migration and widening with time.

The in situ development of Alberta's oil sands has provided the incentive for highly detailed facies and architectural interpretations of the oil-bearing units. Placement of injection and extraction wells for the SAG-D process (steam-assisted gravity drainage) requires a detailed and precise knowledge of the location and extent of oil-rich units, and given the shallow depth of the reservoirs, 3-D seismic is the perfect tool for developing maps and sections that highlight this information. A recent study by Hubbard et al. (2011) is one of the few for which details have been made publicly available. Seismic resolution in this study is better than 5 m. Figure 4.30 illustrates the complex of tidally-influenced point bar deposits and abandoned channels that comprise this unit. Integrating the seismic data with the

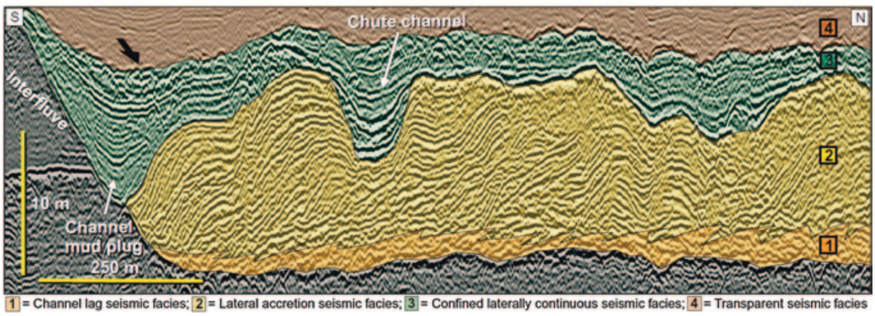


Fig. 4.25 High-resolution sparker-seismic dip-oriented cross-section through a point bar, late Cenozoic fluvial deposits of the Gulf of Thailand. Note convex-up lateral-accretion surfaces, chute channels, and differential compaction overt the channel plug at left, suggesting the predominance of a muddy facies (Reijenstein et al. 2011, Fig. 12, p. 1977). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

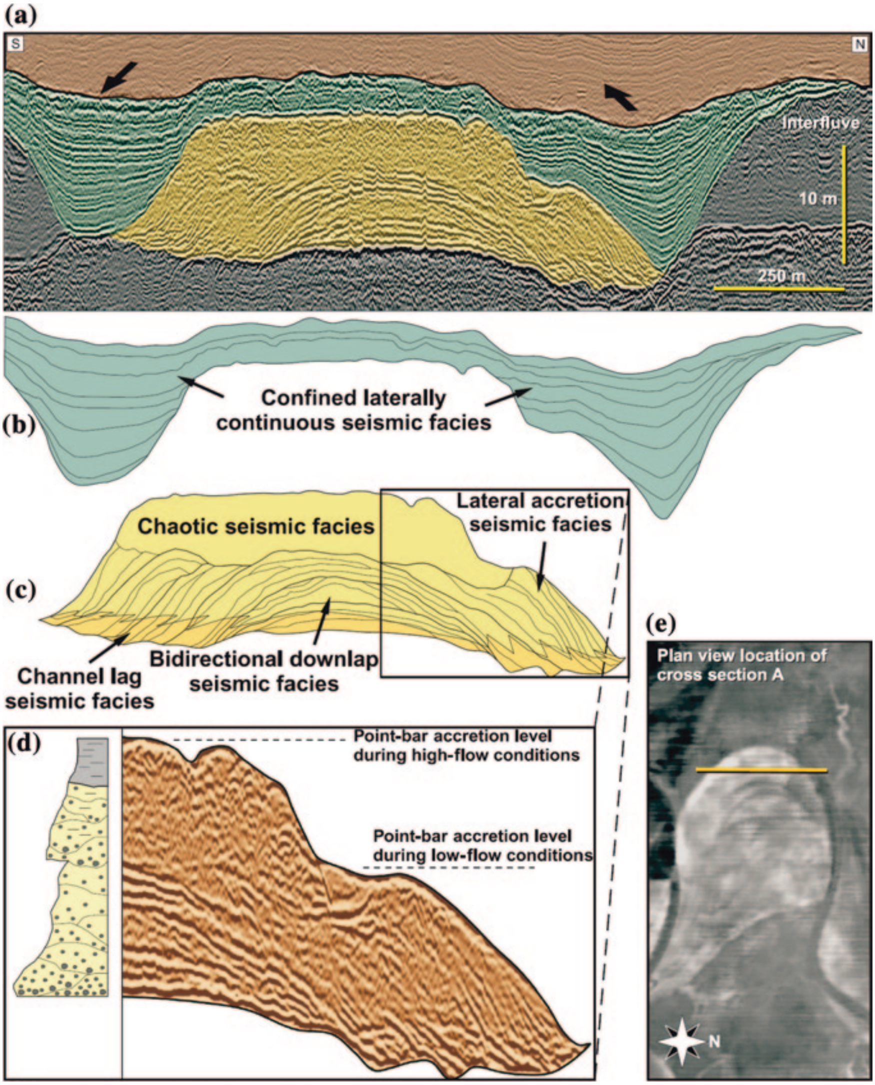


Fig. 4.26 **a** High-resolution sparker-seismic cross-section through a point bar, late Cenozoic fluvial deposits of the Gulf of Thailand. The section is oriented across the front of the bar, as shown in the inset plan (**e**). The two main facies assemblages of the bar are point-bar and channel lag deposits (**c**) and the overlying mud-dominated valley-fill facies (**b**). The margin of the bar is shown in (**d**), together with an interpretation suggesting that the bar was built by discharge at two stage conditions. The vertical profile is speculative (Reijenstein et al. 2011, Fig. 13, p. 1978). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

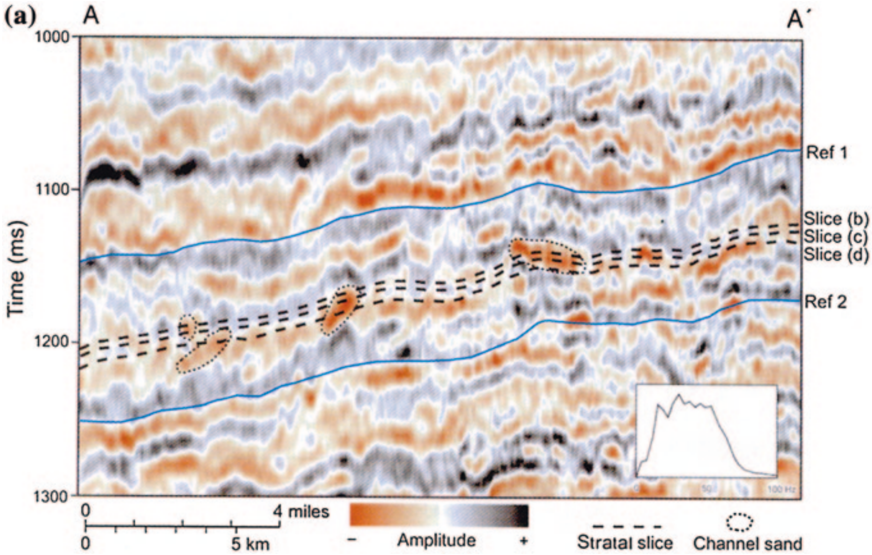


Fig. 4.27 Conventional vertical seismic section through a meandering fluvial channel system (Zeng 2007, Fig. 7a, p. 24)

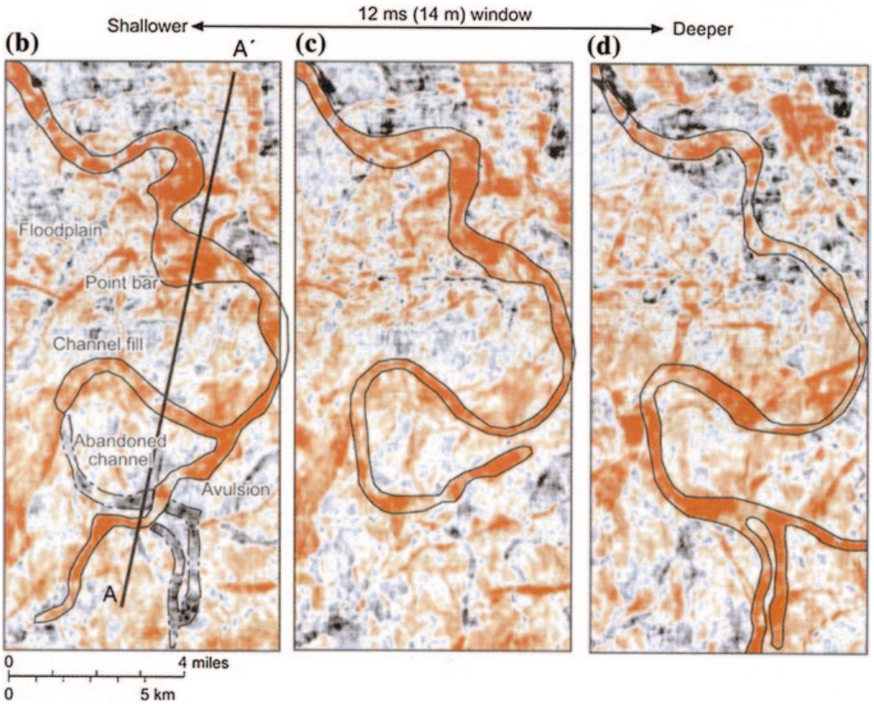


Fig. 4.28 Three stratal slices sampled within a 12-ms window of the section shown in Fig. 4.27 (Zeng 2007, Fig. 7b, p. 24)

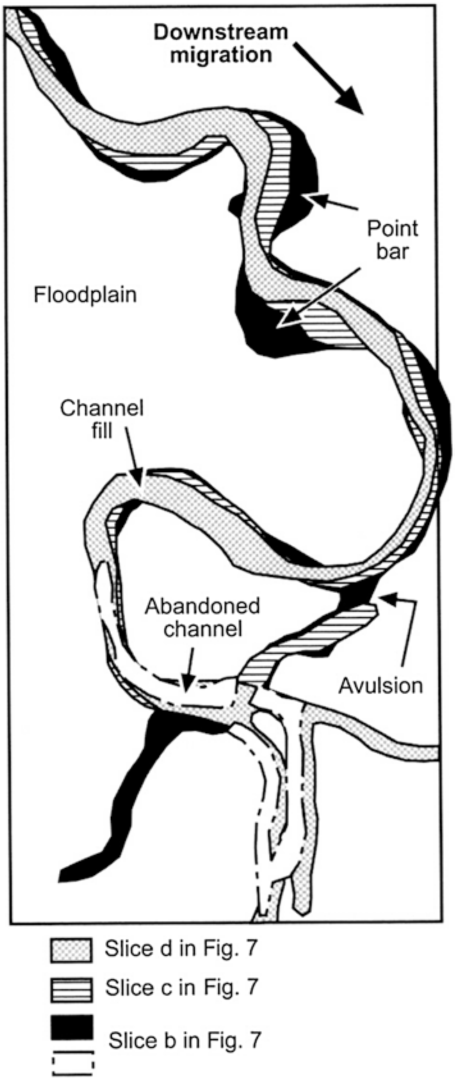


Fig. 4.29 Interpretation of the meanders imaged in Fig. 4.28 (Zeng 2007, Fig. 9, p. 27)

gamma-ray signature yields the map of Fig. 4.30c, where the distribution of the low values, indicating the most sand-rich facies, can be clearly delineated by correlating spot values with the point-bar and channel architecture. Figure 4.31 provides and interpretation of the range of point-bar styles present in this seismic volume.

Time slices through this seismic volume can be interpreted in terms of facies and architectures:

> Scroll patterns associated with point-bar processes are evident, defined by the contrast between interstratified layers dominated by sandstone and siltstone. The overall geometry, composition, and internal stratigraphic architecture of five main depositional elements are

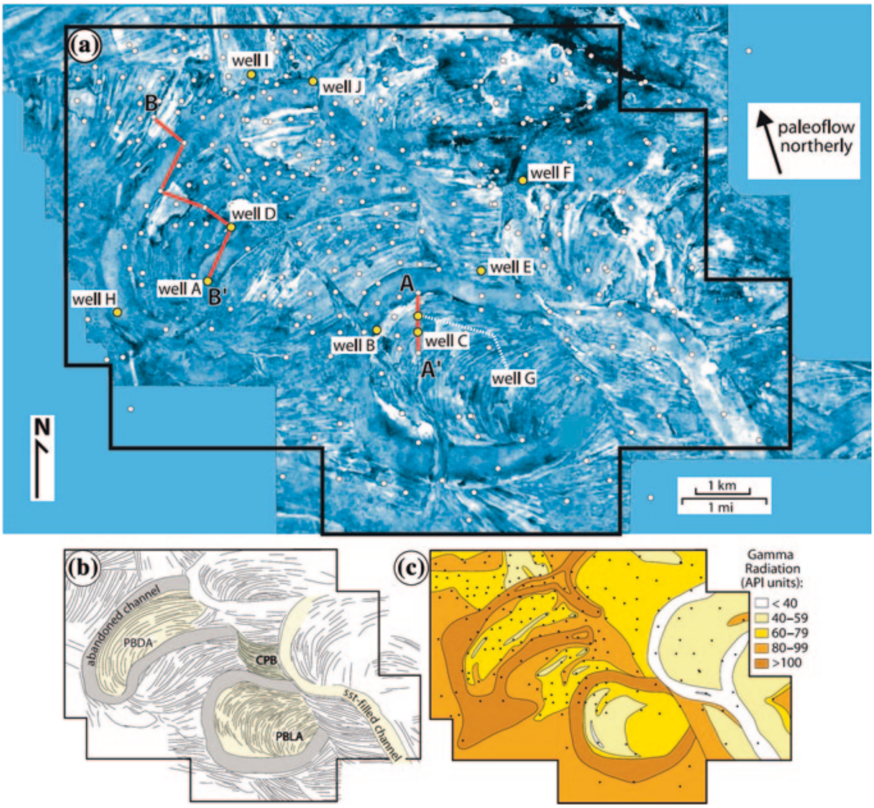


Fig. 4.30 Seismic exploration of the Alberta Oil sands. Mannville Group, south of Fort McMurray, Alberta. **a** Time slice taken at a horizon, 8 ms (approx. 8 m) below a flooding surface at the top of the oil-bearing unit; **b** Interpretation in terms of scroll bars, point bars and channels; **c** Gamma-ray values at the selected horizon. From Hubbard et al. (2011, Fig. 2, p. 1126). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

> described: (1) abandoned channel or oxbow lake fills; (2) point bars associated with lateral accretion (PBLA); (3) point bars associated with downstream, or down-valley, accretion; (4) counter point bars; and (5) sandstone filled channels. (Hubbard et al. 2011, p. 1131)
>
> Abandoned channel reaches are characterized by a relatively homogeneous, moderate seismic amplitude response on seismic time slices. They occur in curvilinear trending bodies, recording the filling of abandoned sinuous channels 400 to 600 m (1312–1969 ft) wide. Drill-core analysis shows that they are primarily composed of a siltstone (Lf4) interval 25 to 35 m (82–115 ft) thick on average, with a cross-bedded sandstone (Lf1) or mudstone clast-rich sandstone (Lf2) layer 2 to 4 m (6.6–13 ft) thick, locally preserved at the base of the succession. Wireline log cross sections demonstrate that abandoned channel fills are asymmetric, gradually tapering out to a zero edge against adjacent point-bar deposits Fill. (Hubbard et al. 2011, p. 1131)

Another example of the use of an integrated data set to characterize the Alberta oil sands was provided by Fustic et al. (2008). In this study, gamma ray and

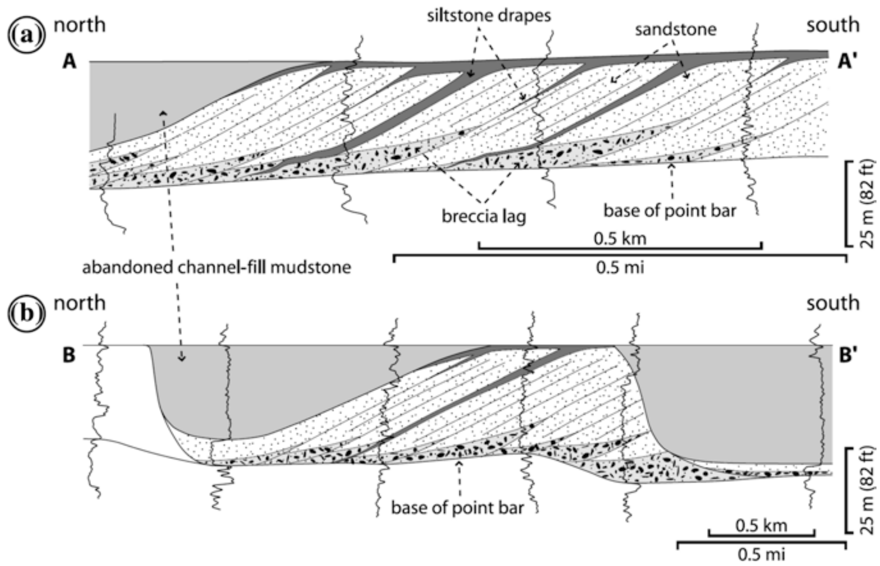


Fig. 4.31 Stratigraphic cross-sections constructed along the lines shown in Fig. 4.30, showing reconstructions of the details of the point-bar architecture, based on seismic and well-log data. From Hubbard et al. (2011, Fig. 6, p. 1133). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

dipmeter data were used in combination with a seismic time slice to map the nesting and orientation of channels and points bars.

A study by Bellman (2010) took the analysis further, by using seismic and wireline-log data to map reservoir facies in detail. Wireline sonic data can be analysed to determine the compressional (P) and shear (S) components, from which shear (mu) and compressibility (lambda) can be calculated. In Fig. 4.32 data so calculated are plotted separately for sand and shale facies, as determined by correlation with core. The results clearly demonstrate that wireline data can be used to discriminate sedimentary facies. Shear and compressibility can also be calculated from seismic data which, of course, can provide this information for every point within a seismic volume instead of just a sample down a drill hole. Figure 4.33 shows a seismic section that has processed using this inversion method to develop a facies analysis. Gamma ray logs on this section were drilled after the seismic facies interpretation had been constructed, and facies predictions made from this analysis agree with those made from the gamma-ray logs over 75 % of the total footage drilled.

Steam injection turns bitumen from a solid to a liquid. The unaltered bitumen can transmit a shear wave, whereas the heated, fluid cannot, so this technique, when run repeatedly during production, can be used to follow the process of bitumen drainage of a reservoir volume.

Figures 4.34 and 4.35 illustrate other examples of fluvial architectural features within the Freshwater Molasse of the Alpine foreland basin in Switzerland. In the

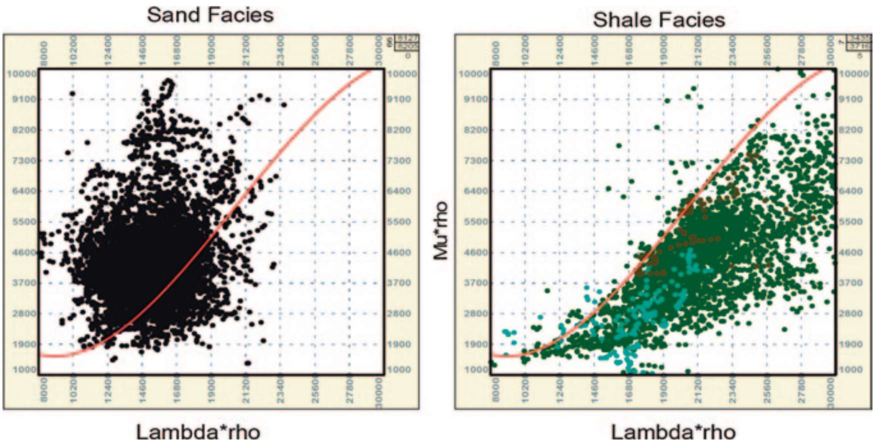


Fig. 4.32 Cross-plot of computed well-logs from 85 wells, with facies separated on the basis of core analysis. Bellman (2010, Fig. 2). AAPG © 2010. Reprinted by permission of the AAPG whose permission is required for further use

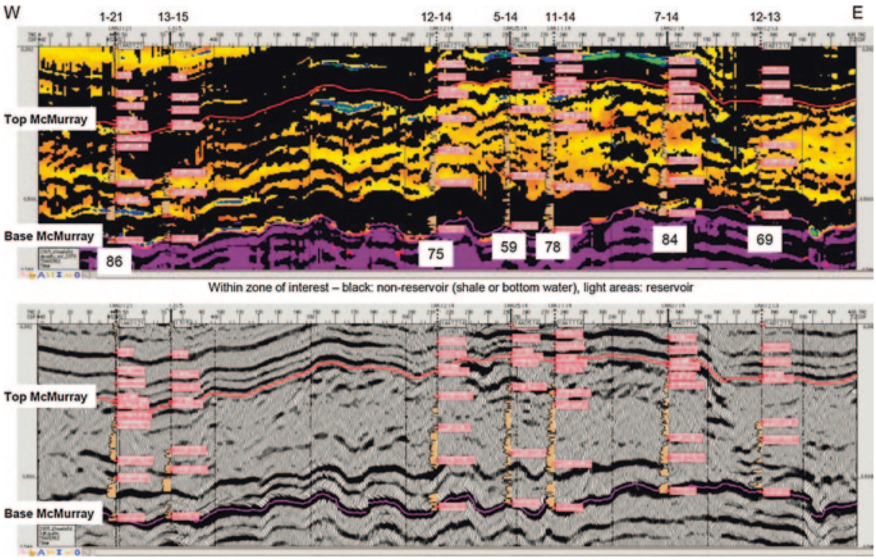


Fig. 4.33 A seismic profile (below) and a derived profile in which compressibility and shear data have been interpreted in terms of reservoir and non-reservoir facies. *Black* shale or bottom water-filled units; *yellow* bitumen reservoir; *blue* wet reservoir; *green* gas reservoir. Also shown are gamma ray logs, drilled after the seismic survey and analysis had been completed. From Bellman (2010, Fig. 3). AAPG © 2010. Reprinted by permission of the AAPG whose permission is required for further use

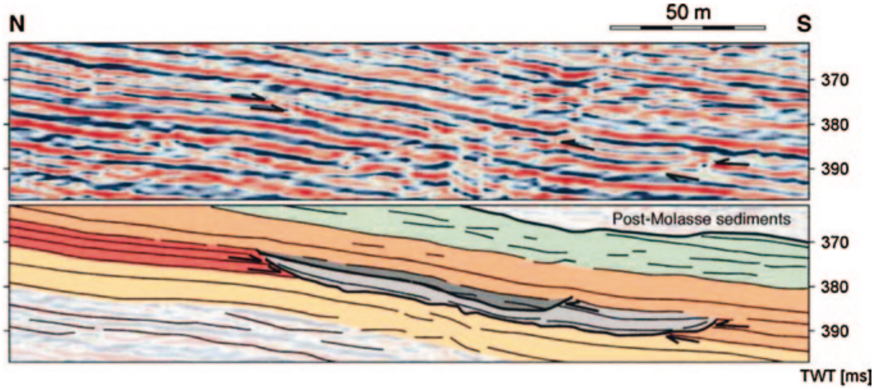


Fig. 4.34 Seismic image (uninterpreted and interpreted) of nested channels from the Freshwater Molasse of the Swiss Alpine foreland basin. From Morend et al. (2002, Fig. 7, p. 255)

uninterpreted line shown in Fig. 4.34 half-arrows are used to point to truncation of reflections by a scoop-shaped feature. This is interpreted as a pair of nested channels, the wider one being about 150 m across. Figure 4.35 illustrates an example from the same stratigraphic unit of an incised valley about 50 m deep and at least 350 m wide sampled along two survey lines about 100 m apart. The valley fill can be subdivided into two distinct seismic facies packages. The lower part of the valley fill shows discontinuous, low-amplitude reflections, and it is suggested that this may be an indication of valley-margin slumps, which have generated a chaotic reflection character. The upper part of the valley fill would then represent an undeformed valley fill, probably deposited during a rise in local base level.

Hentz and Zeng (2003) and Zeng and Hentz (2004) completed a sequence analysis of the Miocene and lowermost Pliocene continental slope and shelf deposits of part of the Louisiana offshore using seismic data tied to wireline and biostratigraphic data from offshore wells. Their study revealed many depositional features that offered the possibility of seismic-geomorphic analysis. Wood (2007) followed up with a detailed analysis of fluvial features in this very large data set that contained thirty-seven identifiable surfaces for seiscrop mapping in the more than 3 km of section (Fig. 4.36). Wood turned to the literature on fluvial geomorphology containing quantifiable data on channel widths, depths, meander geometries, the ratios that relate these features to each other, and the relationships of this information to fluvial styles and lithologies (the data discussed in Chap. 2 of this book). She identified three classes of incised fluvial system (figure numbers in this quote have been modified to the figure numbers used in this book):

> Three types of channelized systems are visible in seismic geomorphologic slices taken from the study area. On the basis of their geometry, Class 1 systems are interpreted as large, aggradational fluvial systems with large meander-belt widths, high sinuosities, and large meander-arc heights (Fig. 4.37). These systems are both transportive and depositional, forming extensive floodplains, complete with large, abandoned oxbow lakes (Fig. 4.37). In contrast, Class 2 systems are interpreted as bypass fluvial valleys,

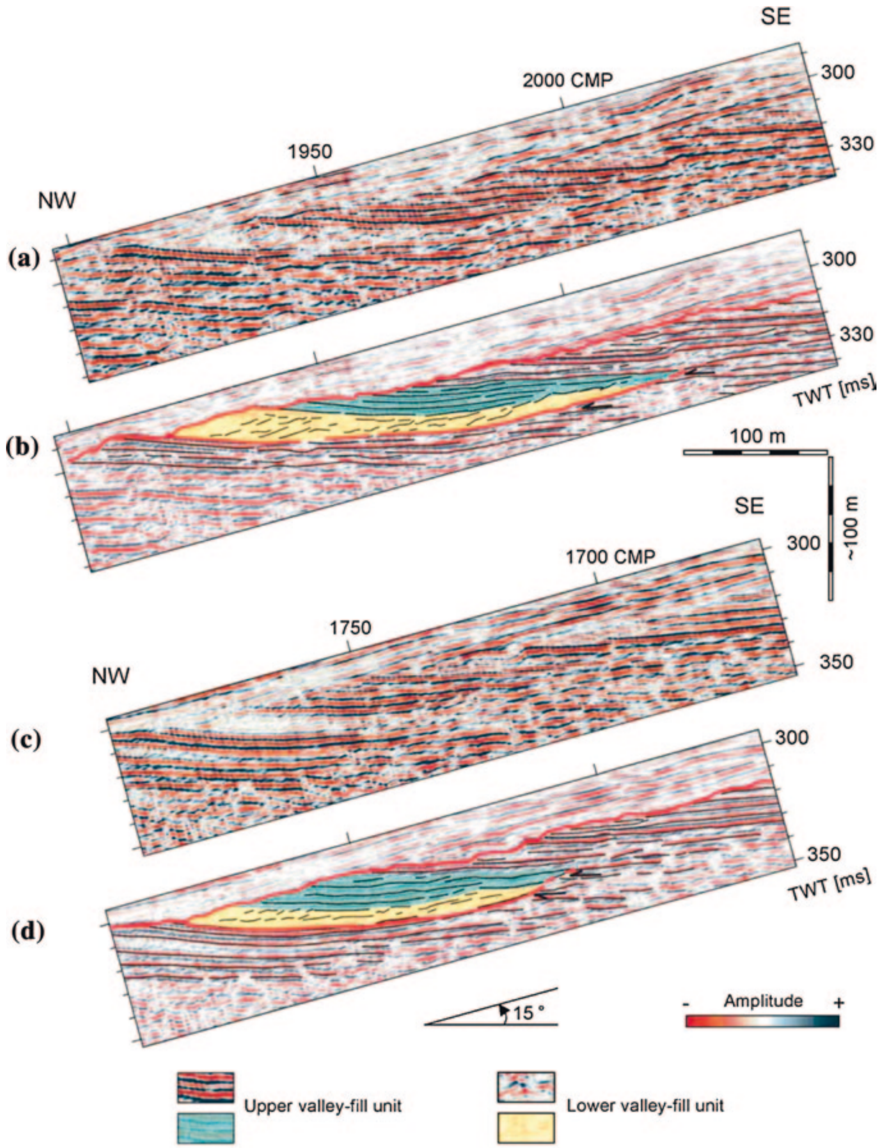


Fig. 4.35 Seismic image of an incised valley fill, overlain unconformably by Pleistocene glacial deposits. Freshwater Molasse of the Swiss Alpine foreland basin. From Morend et al. (2002, Fig. 9, p. 258)

showing significantly lower sinuosity, well-defined edges within the study area, and small meander-arc heights (Fig. 4.38). These systems are associated mainly with maximum lowstand surface sequence boundaries, showing significant incision. They are often filled with a multitude of individual channel types and are dominantly filled with a sandy lithology (Fig. 4.39). Class 3 systems make up a wide variety of architectural elements

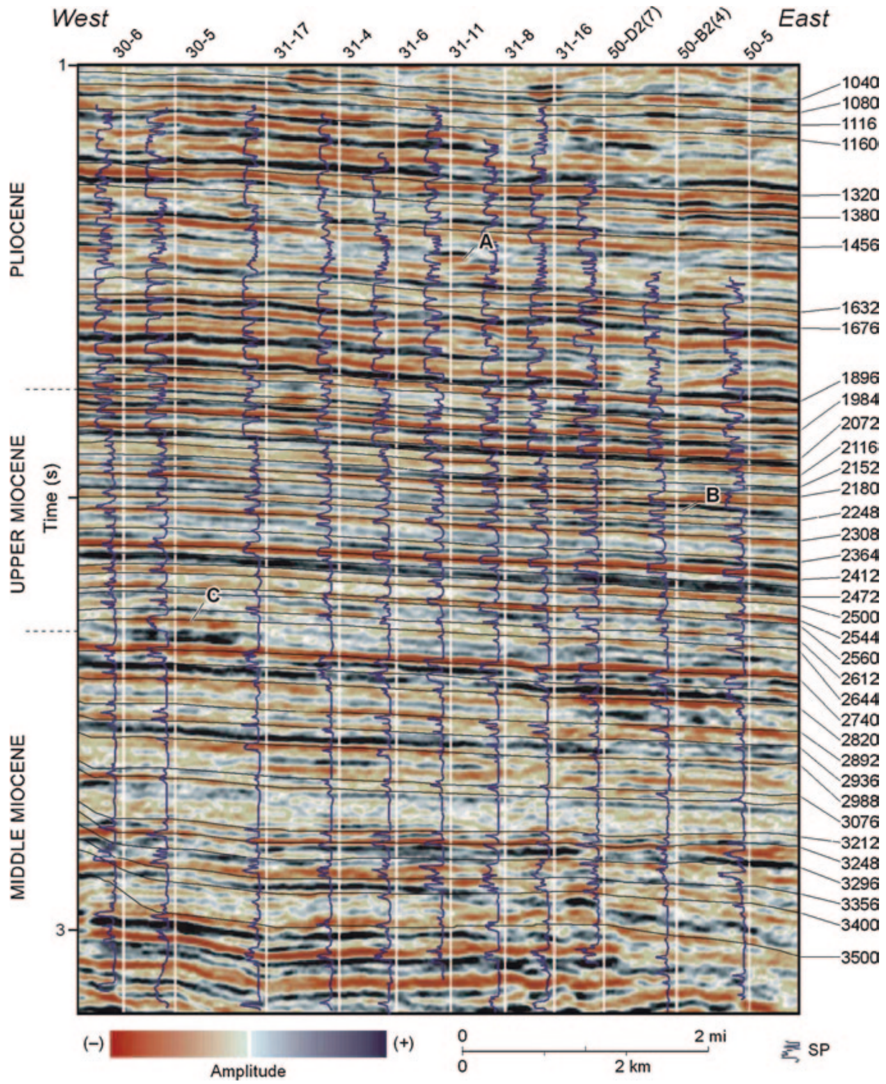


Fig. 4.36 Cross-section through the Miocene-Pliocene seismic volume from offshore Louisiana, showing spontaneous logs, as analysed by Wood (2007, Fig. 5, p. 717)

> in the fluvial–deltaic coastal-plain system, including distributary channels, tidal creeks, and interdistributary drainages (Fig. 4.40). They form narrow meander belts with highly sinuous, often crenulated channels and, more infrequently, nearly anastomosing. They frequently show mappable meander-migration architecture. These channels most likely drain limited areas in and around the coastal plain and shoreline or are distributive feeders to small deltas immediately south of the study area. (Wood 2007, p. 723)

The lithologic composition of the channels can be determined from gamma ray logs, using a method devised by Zeng and Hentz (2004). This shows that class 3

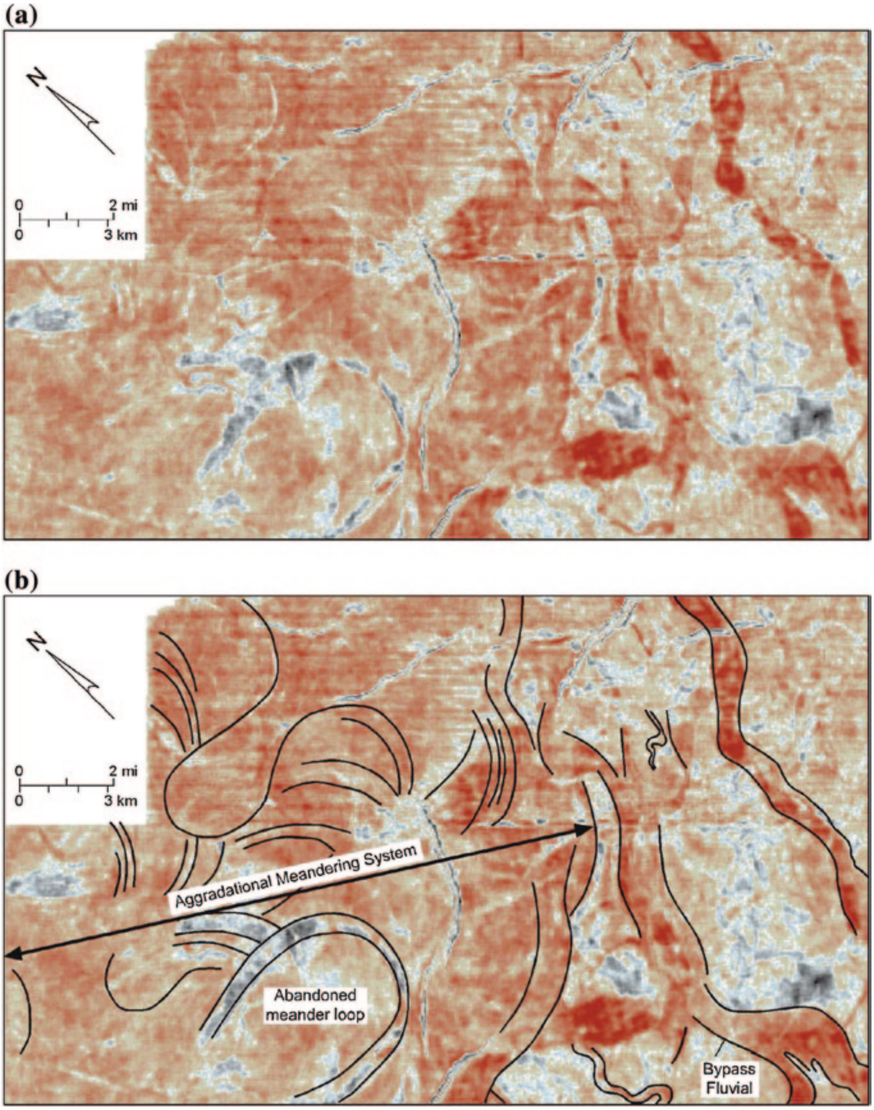


Fig. 4.37 Seiscrop section in the Miocene of offshore Louisiana showing a large Class 1 aggradational meandering fluvial system with some low-amplitude abandoned meander loops. This Class 1 system is impinging upon an older Class 2 lowstand bypass fluvial system to the east-southeast. (Wood 2007, Fig. 7, p. 710)

channels are of suspended load type (using the classification of Schumm (1977); e.g., see Fig. 2.23), whereas classes 1 and 2 ranged between bedload, mixed load and suspended-load type. As Wood (2007, pp. 728–729) noted, the ability to map fluvial systems and predict their lithofacies could be a very useful component of an exploration program.

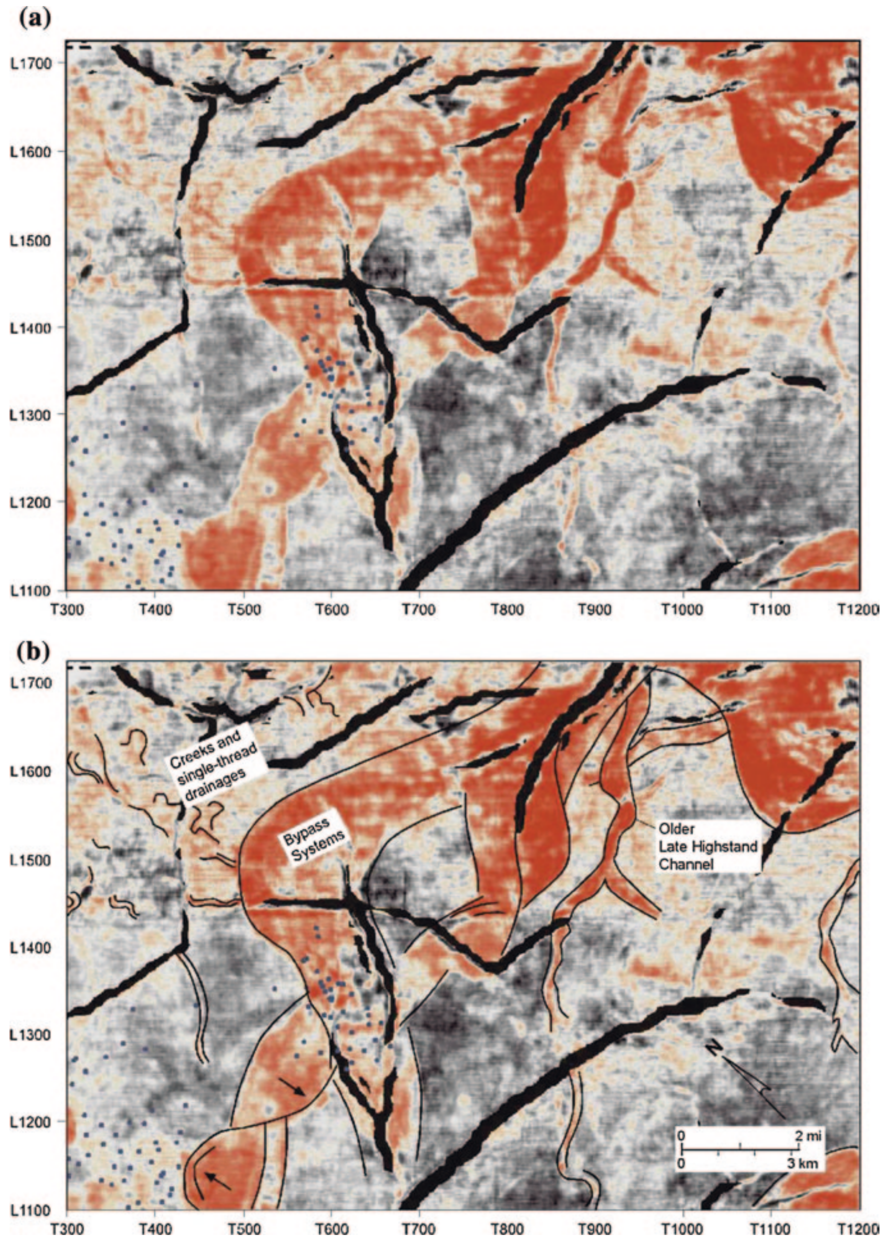


Fig. 4.38 Seiscrop section in the Miocene of offshore Louisiana showing a large Class 2 bypass system incising into older late highstand distributary channels that bifurcate to the southwest. *Blue dots* represent wells penetrating the interval. Smaller Class 3 creeks and single thread channels are draining the surrounding coastal plain to the north. *Thick black lines* represent major faults across the area (Wood 2007, Fig. 8, p. 721)

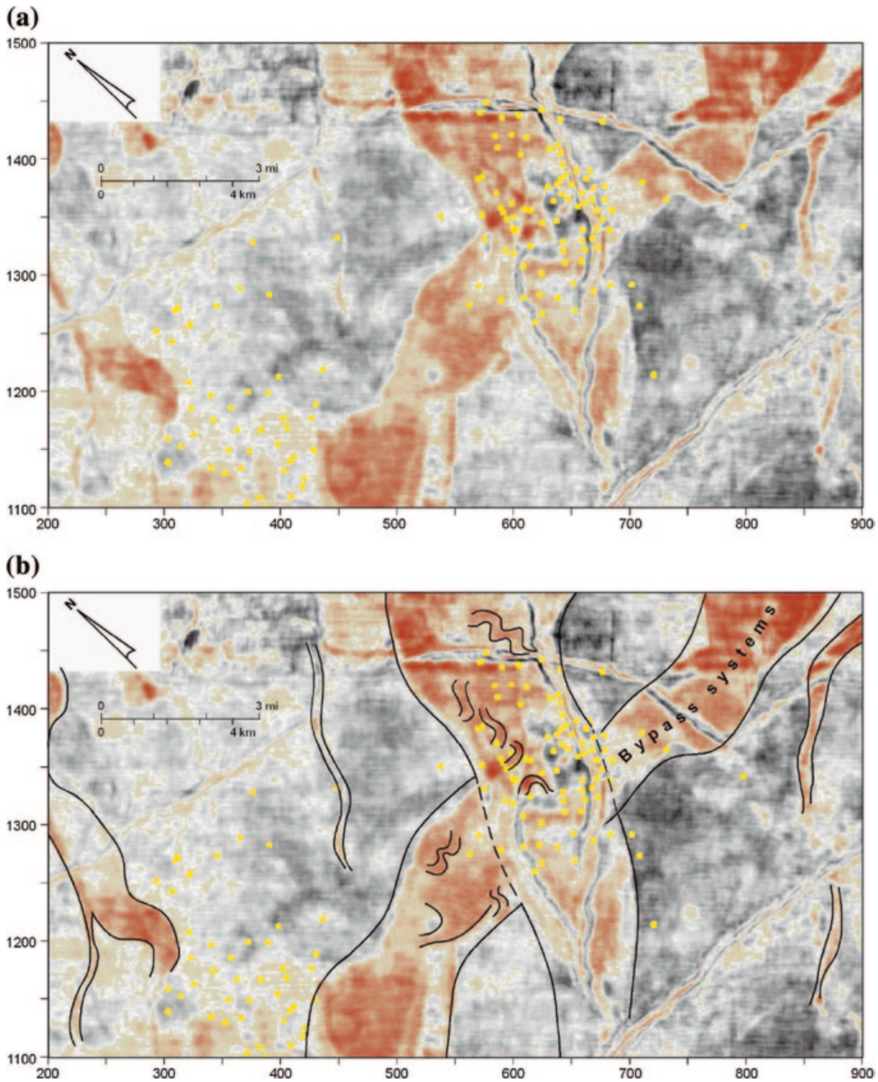


Fig. 4.39 Seiscrop section in the Miocene of offshore Louisiana showing a large Class 2 bypass valley system with internal channels several hundreds of meters wide. *Yellow dots* are well locations. Several *thick black lines* mark major faults across the area (Wood 2007, Fig. 10, p. 724)

Miall (2002) analysed a series of closely-spaced time slices of the Pleistocene fluvial succession beneath the Malay Basin. One of these is illustrated in Fig. 4.41, and its interpretation in Fig. 4.42. One of the most interesting features of this data set is the successive sampling, at different levels, of a valley fill with a meandering system at its base, and an incised and gullied tributary, all of which are clearly visible in the eastern part of the 196 ms image. The incised tributary is somewhat

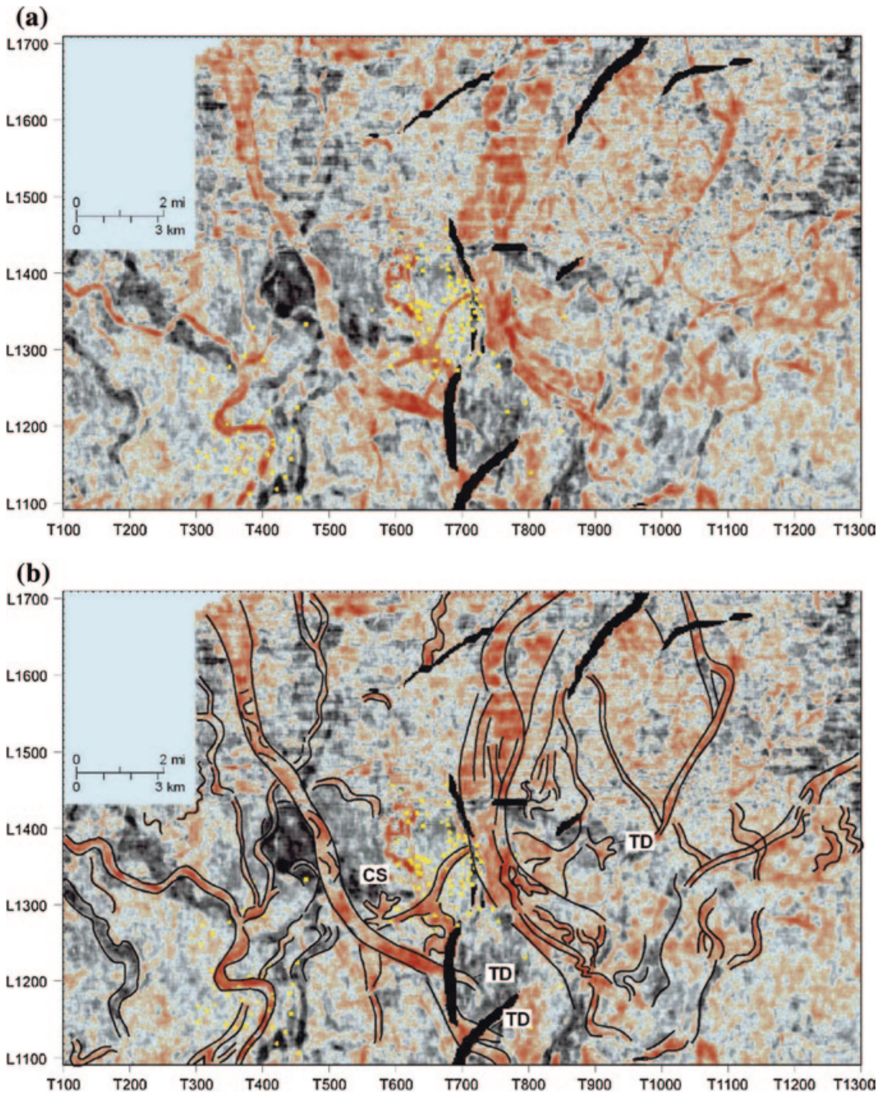


Fig. 4.40 Seiscrop section in the Miocene of offshore Louisiana showing multiple Class 3 creeks and distributary channels cutting the older highstand coastal plain, overlain by at least two distinctly larger (2–3 km wide) Class 3 systems with multiple internal channels and accretionary architectures. Several of these channels break into terminal distributary lobes (TD), and it is possible that a few crevasse splays (CS) are identifiable. *Yellow dots* are well locations. Several *thick black lines* mark major faults across the area (Wood 2007, Fig. 11, p. 725)

wider than in the underlying plot (typically up to 600 m, vs. the 300 m width in the 208 ms slice), suggesting a V-shaped cross section, as would be expected for an incised valley. Assuming a simple triangular cross section, this corresponds to

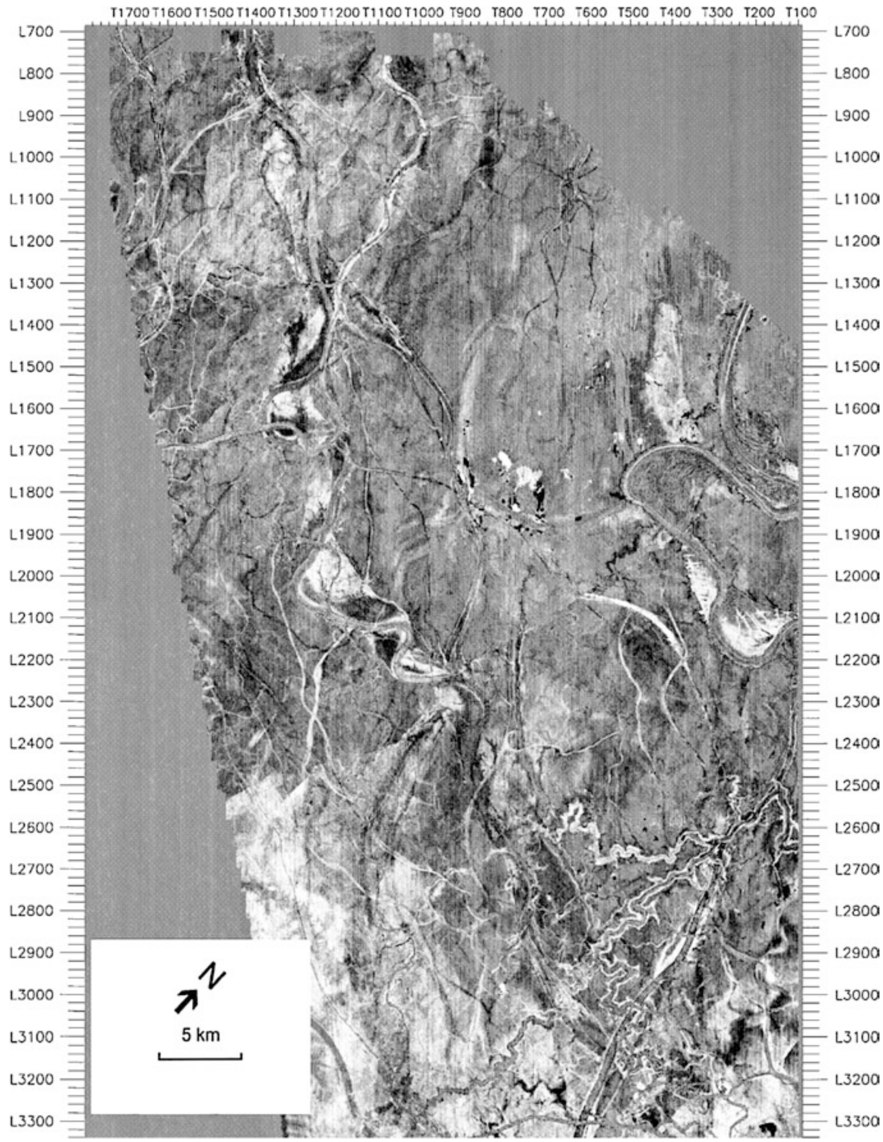


Fig. 4.41 A seismic time-slice image of Pleistocene fluvial deposits, 196 ms interval, Malay Basin. Location of this map and of the data provided in Figs. 4.42 and 4.43 is given in Fig. 7.11, project #2. From Miall (2002, Fig. 2, p. 1205)

valley sides sloping at about 4°. Figure 4.43 is a perspective view of the eastern corner of the five successive time slices, superimposed to show how the meandering river and its tributaries change from one level to another.

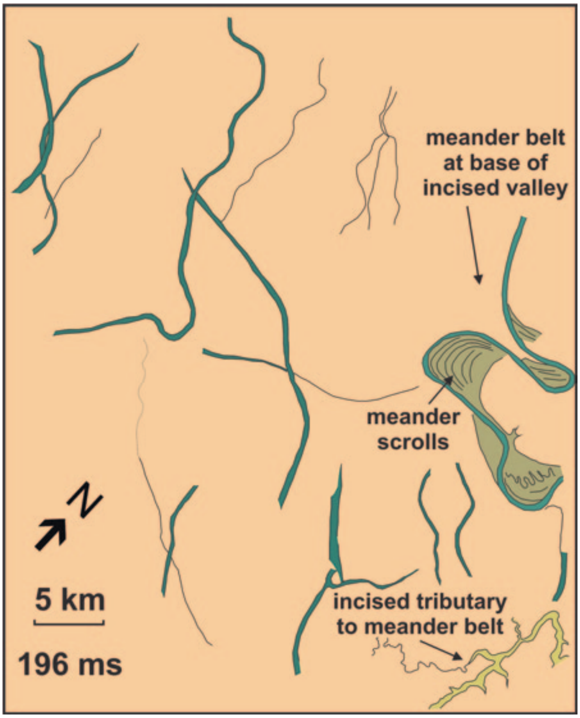


Fig. 4.42 Interpretation of the 196 time slice, Malay Basin, from Miall (2002, Fig. 4, p. 1206)

Numerous other minor channels are visible, mostly of low sinuosity. Some of these elements appear in more than one time slice, which is a well-known "shadow" effect of seismic data.

Given adequate seismic data and well control, the imaging of incised valleys and other features may make a substantial contribution to the development of a sequence interpretation of a stratigraphic succession. An example was described in detail by Maynard et al. (2010), from a Cretaceous heavy-oil field in Alberta. The stratigraphy in this data set was subdivided based on flooding surfaces recognized and correlated in seismic and well data. Facies were identified and classified on the basis of core analysis, and tied to the well and seismic data, yielding a detailed regional paleogeographic interpretation (not shown here). A highlight of the analysis is the recognition of a series of intersecting incised valley systems (Figs. 4.44, 4.45, 4.46). Facies analysis indicates that these are filled with fluvial to estuarine facies.

4.3.2 Surveillance Methods

As applied to the petroleum industry, the term "*surveillance methods*" means methods of measurement and observation on a producing field that work to monitor reservoir performance, including drainage patterns, changes in pressure,

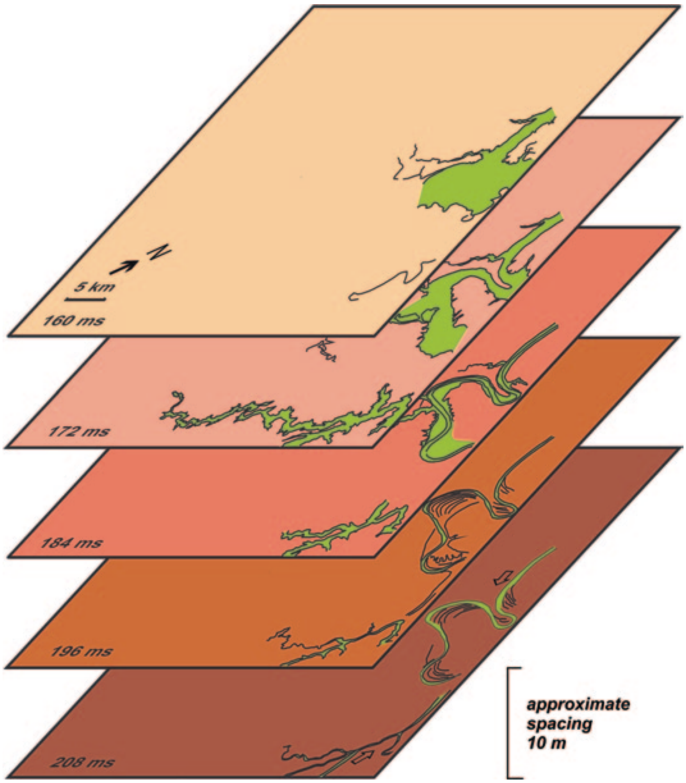


Fig. 4.43 A superimposition of five of the time slices drawn to emphasize the vertical succession of features along the east side of the project area in the Malay Basin. The major feature of shown is a valley-fill, with a tributary entering from the south. Note the meandering channel with scroll bars in the lower two images and the major V-shaped tributary valley along the southeast edge of the project area, which becomes wider as it is intersected in successively younger and higher time slices. The valley fill above the main meandering system is visible in the 172 and 160 slices (Miall 2002, Fig. 10, p. 1212)

water-cut, etc. A wide range of methods has emerged over the last couple of decades to provide this type of production information, the value of which is that it can assist the production engineers, in real time, in adjustments or modifications to production models in order to maximize production efficiency.

Nearly two decades ago, He et al. (1996) stated: "4D, or time-dependent, seismic reservoir monitoring is an emerging technology that holds great hope as an oil-production management system." (p. 41) "… we are only just beginning to visualize the changes that occur within reservoirs as oil and gas are drained over time, and the changes that we see are surprising us. For example, gravity is often not nearly as efficient at sweep as we thought it was for all these years." (p. 42).

The example they highlighted in this report was the BP/Shell Foenhaven field off the western Shetlands. In the mid-1990s a permanent bottom cable seismic

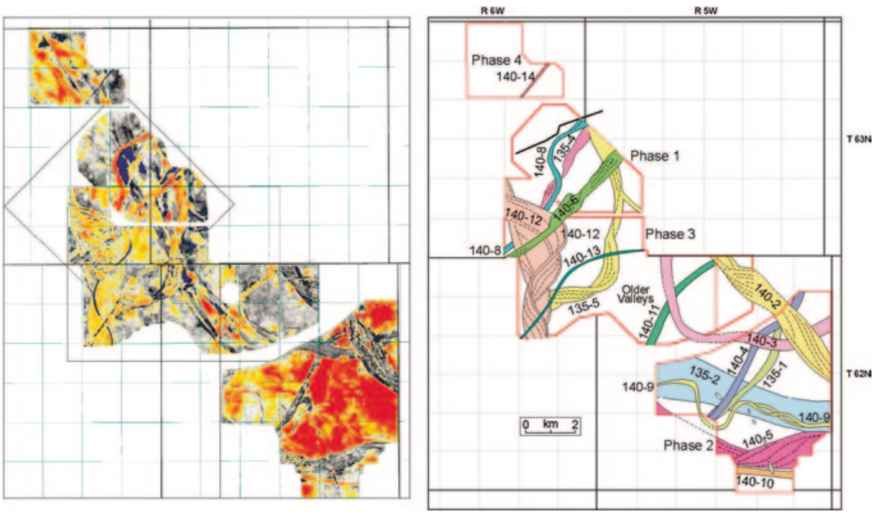


Fig. 4.44 A time slice illustrating a complex of incised valley-fills in the Iron River Field, Alberta. Each square is 1 mile (1.6 km) across (Maynard et al. 2010, Fig. 2, p. 613)

array had been installed over the field. Repeated surveying was then used to "right-size" the volume and rate of production in this deep-water system. As He et al. (1996, p. 42) stated: this type of technology could lead to "improved location and timing of development drilling programs, early verification of the reservoir simulation model, and improved in-fill drilling and injector placement."

> A quantitative reservoir simulation is constructed through use of both inverse and forward seismic models of 4D seismic differences that can then be iteratively recomputed and compared with the initial reservoir simulation. "Dynamic changes in the hydrocarbon reservoir can be monitored and simulated efficiently, and the results can then be used to understand and predict drainage occurring during production. New wells can then be placed to maximize the lives of oil and gas fields. To achieve the highest hydrocarbon recovery rate possible". (He et al. 1996, p. 43)

Amongst the invaluable bodies of information generated by the Shell survey data was the revelation that gravity had not been as important a driver of drainage as had been thought. 4D data showed a complex pattern, probably reflecting water fingering, most likely controlled by sand quality variations. Inefficient gravitational sweep left many high-amplitude, low impedance zones downdip. In their example, the identification of a bypassed zone led to the placement of a horizontal well that then produced more than 1 million additional barrels of oil.

Smalley et al. (1996) provided an excellent summary of surveillance methods. Here is their summary of methods—a "reservoir compartmentalization toolkit" being used at that time:

3D seismic interpretation;
Pressure data;
Oil geochemistry (molecular parameters, GC fingerprinting);

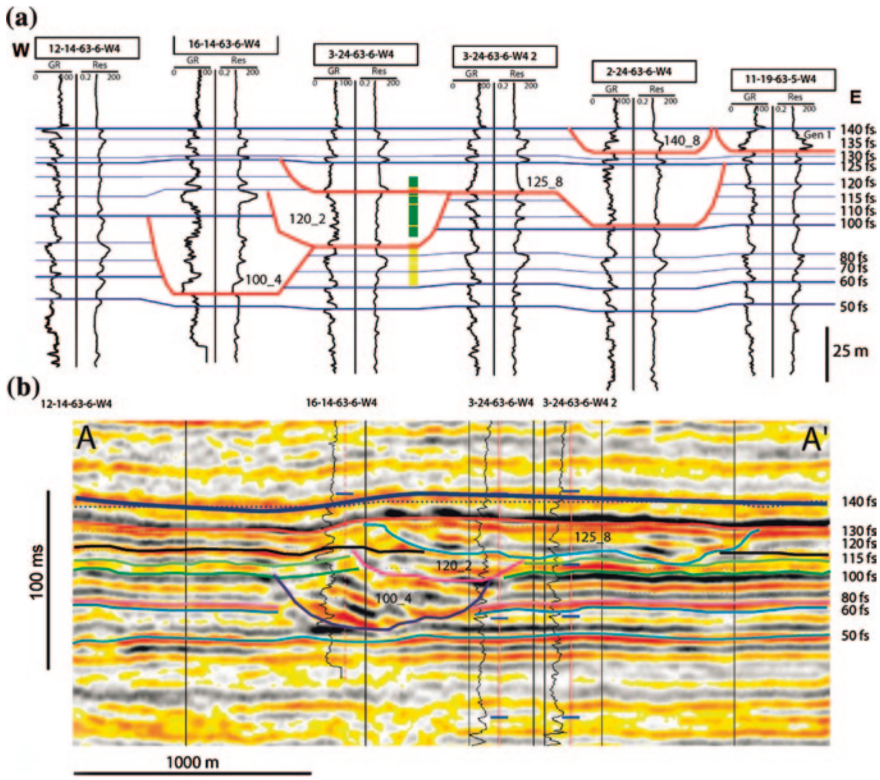


Fig. 4.45 **a** Well correlation line shown in Fig. 4.44. Fluvial and estuarine sandstones and shales valley fills incised into shallow marine shoreface deposits. The geometries are complex and without the seismic it would be difficult to separate some of the valleys. Core logs are are colored according to interpreted facies: *yellow* shallow marine, *orange* fluvial sandstones, and *green* fine-grained nonmarine facies. **b** A seismic section through part of the same section. Note the parallel reflections of the shoreface deposits (high to moderate amplitude continuous parallel seismic facies) that are truncated by the incised valleys. Valley 100-4 shows moderate to high amplitude semi-continuous shingled seismic facies interpreted as lateral accretion surfaces. Valley 125-8 shows high to moderate amplitude discontinuous seismic facies (Maynard et al. 2010, Fig. 7, p. 619)

Oil PVT properties;
Well test analysis;
Fault seal analysis;
Formation water composition;
Residual salt analysis (RSA);
Reservoir heterogeneity modeling;
High-resolution stratigraphy;

To be used in fields with some production history:

Tracer data;
Pressure decline analysis;
4D seismic.

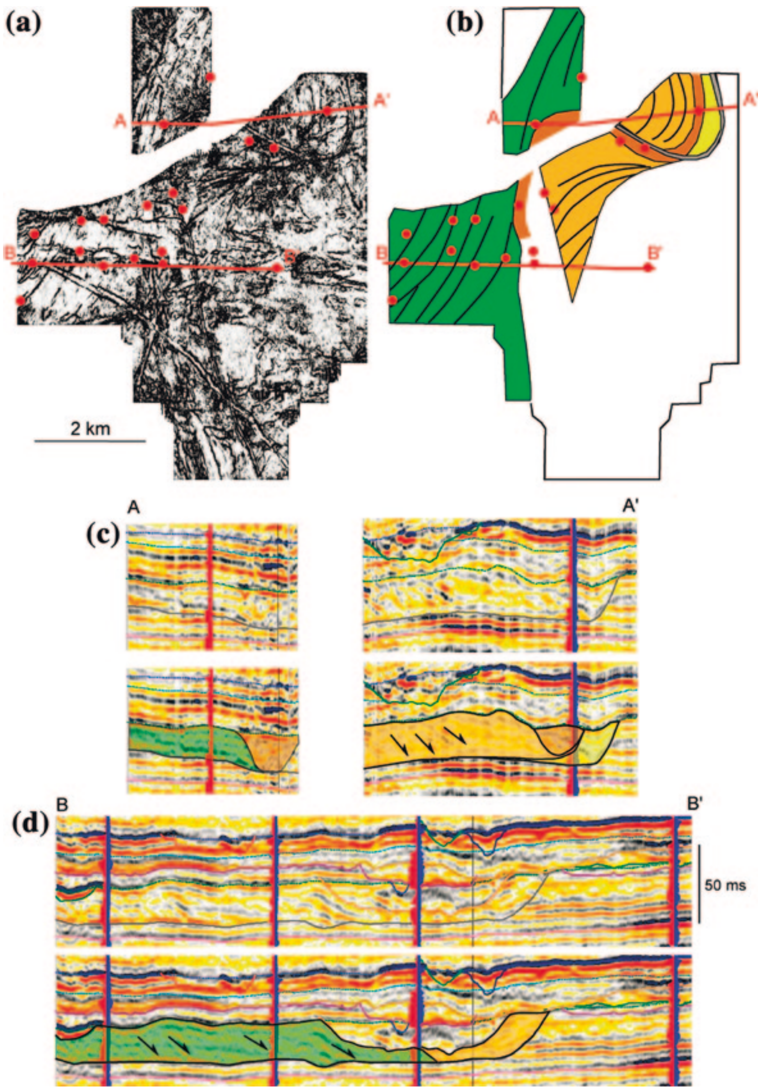


Fig. 4.46 **a** Uninterpreted and **b** interpreted amplitude time slice from just below 100 flooding surface showing the generations of channel fills for the 100_5 valley fill. *Green*, *orange*, and *yellow* are different point bars. *Brown* is the abandoned channel fill at the end of the orange point bar. *Grey* is a small channel that is barely at detection level for the seismic that was the last event recorded in the valley fill. **c** Cross sections A–A′ and **d** B–B′ showing sequence boundaries (*solid lines*) and flooding surfaces (*dashed lines*). Moderate to high amplitude semi-continuous shingled seismic facies fills these incisions cut into high to moderate amplitude continuous seismic facies. Lower version of each shows channel elements from 100_5 valley fill using same colors as in b. Well logs show GR (*red*) and resistivity (*blue*) curves. Note the presence of valley 100_8 at the *left* side of B–B′ just below the 100 flooding surface (Maynard et al. 2010, Fig. 11, p. 626)

For example, pressure-depth plots could show two wells to be on a different curve indicating compartmentalization, which could be attributed to depositional architecture or to faulting. Oil molecular maturity and gas chromatography data could show the same compartmentalization. Residual salt analysis (RSA) is another tool. The method is to recover dissolved solids that precipitate out in core during storage. This provides a $^{87}Sr/^{86}Sr$ ratio from original pore waters, which experience shows evolve differently in separate reservoir compartments.

As they stated, "integration is the key." "Unfortunately, compartmentalization is notoriously difficult to identify during the early history of a field, often only coming to light as new wells are drilled and dynamic data accumulate during field development. This, of course, may be too late! The discovery that a greater number of wells will be needed, or that surface facilities are inappropriately sized or located, may already have compromised the economics of the project" (Smalley et al. 1996, p. 163). They noted that reservoir compartmentalization can be constrained and quantified during reservoir appraisal, even in the absence of dynamic production data.

Another useful treatment of the topic was provided by Villalba et al. (2001), They dealt with the following techniques:

> Pressure-depth plots reveal unconnected reservoir units.
> Measurement of API gravity (This changes with time, with lightest gravity oils produced first). Again, differences indicate conection or lack of between flow units.
> Oil fingerprinting using gas chromatography to identify biomarkers.

Mezghani et al. (2004) discussed history matching and the use of 4-D seismic. Compressional/shear impedance data from 3-D surveys are used to calculate petrophysical properties and refine production models in successive 3-D surveys. "Most of the 4-D seismic studies show it is especially possible to follow gas-oil contact displacement because the latter makes appear important reflexions. Well, gas apparition causes an important decrease of seismic velocities leading to an impedance variation [sic]." (Mexghani et al. 2004, p. 5)

Andersen et al. (2006) developed a geological model from seismic inversion data, and used that to construct a flow model and compare it with 4D seismic data.

> Cross plots of the changes in elastic parameters from 4D seismic can be used to classify 4D effects into saturation related or pressure related changes. The proposed workflow is combining both 3D and 4D elastic inversion data to classify lithology. In the case of the 4D seismic, it is assumed that some of the observed production effects mainly can be related to sand facies only. By utilising this, it is possible to achieve a sand probability cube that has highest probabilities when both 3D and 4D seismic are utilised. (Andersen et al. 2004, p. 1)

Kaufman et al. (2000) used a different technique. Sixty oils from Burgan field, Kuwait were analyzed using oil fingerprinting, to see if individual reservoirs had unique compositions and whether oil from different horizons was mixing during production. Bulk properties of oils can be used to characterize them, including oil gravity, gas/oil ratio, and bubble-point data.

Molecular composition is determined by whole-gas chromatography and is the most sensitive way to characterize oil for this purpose. Differences are initially related to thermal maturity. Cluster analysis of chromatography data revealed three groups of oils each corresponding to a different reservoir. However, oil fingerprinting and pressure-depth plots for the three Burgan sands suggested good fluid communication between these units. Some of this could be due to production cross-flow. The 50-year production history of this field might have allowed some leakage between reservoir zones, so the fingerprinting data cannot be relied on completely to indicate what the primary flow paths and connectivities were. The authors recommended that fingerprinting be done on samples taken before production begins.

Westrich et al. (1999) showed that gas:oil ratio (GOR), AIP gravity, and pressure–volume-temperature (PVT) data demonstrate different values for a later well brought on stream along the flank of the field, suggesting a permeability barrier. Chemical differences between oils demonstrated a permeability barrier within another of the reservoirs. This could be structural or stratigraphic in origin (Fig. 4.47).

> Compositional gradients in continuous reservoirs, mixing processes, and the underlying causes and controls of observed chemical differences need to be understood for proper interpretation. (Westrich et al. 1999, p. 518)

Another example of the use of geochemical and production methods was described by Refunjol and Lake (1999). They used six different gas tracers over

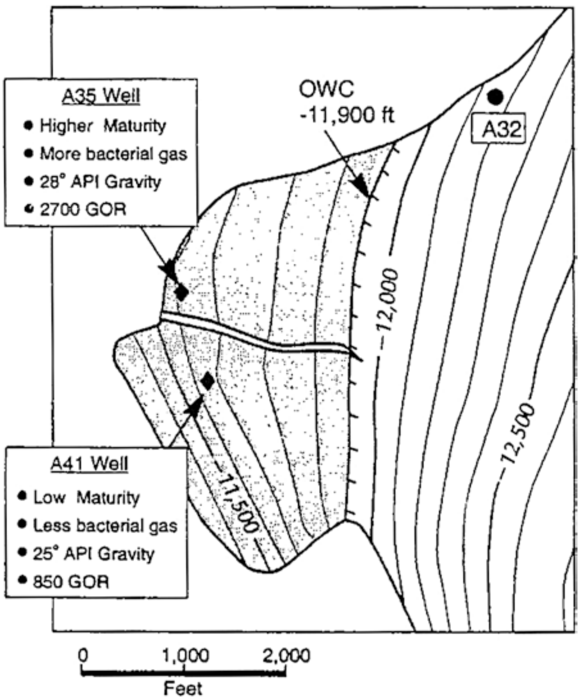


Fig. 4.47 Geochemical differences in the oil and gas in two wells in the Bullwinkle field, Gulf of Mexico. The differences are attributed to either a structural or stratigraphic barrier causing reservoir compartmentalization (Westrich et al. 1999, Fig. 7, p. 517)

a four-year period to test injector-producer communication in the North Buck Draw field, Wyoming, a field producing from a complex valley-fill succession. Tracer recoveries are shown by the arrows in Fig. 4.48. No recoveries were made from some of the injectors (e.g., 22–17, 22–20 and 22–31) on the east side of the reservoir.

> Tracer or tagged gas production at producers in the west, some at early and some at late break-through times, indicated no sealing geological barriers are present. Conversely, the lack of response to tracers injected in wells on the east side of the reservoir indicates no communication or, at least, less communication between these wells and the rest of the reservoir.

The second method used in this study was to compile monthly injection-production rate data between each injector-producer pair. The data were converted to ranks, and the values compared over periods of months. Given that the

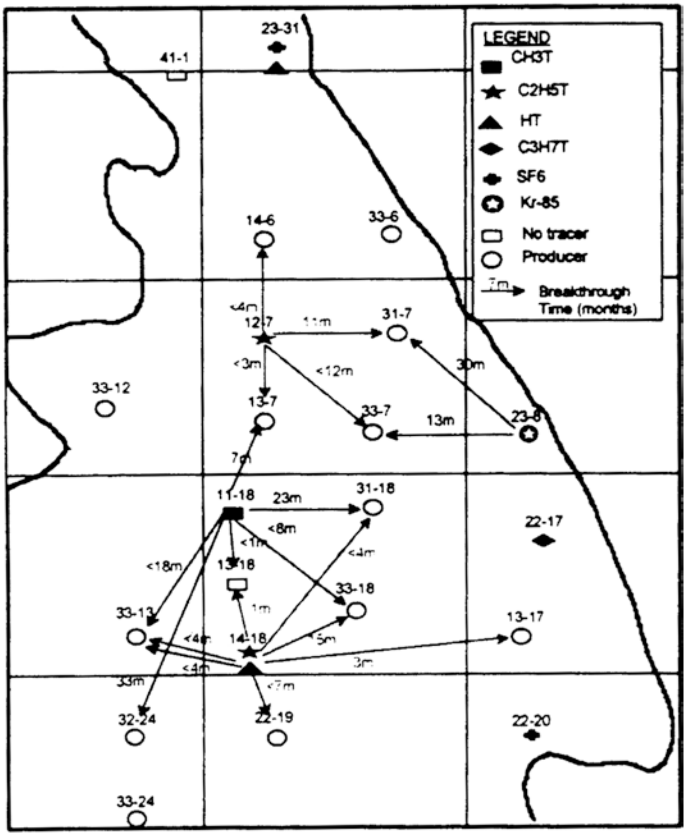


Fig. 4.48 Tracer response pattern. The six codes at the top of the legend refer to six different tracers used. *Arrows* indicate tracer recoveries between injector and producer (Refunjol and Lake 1999, Fig. 2, p. 212)

response from injector to producer is a significant time lag, calculations of correlations between injector and producer were lagged from zero to a lag corresponding to half of the total number of months. Maximum correlation values were derived when the lag time was set at 13 months. It was found that when the orientations of injector-producer well pairs was compared, highest correlation values were obtained when the orientation direction between the wells was in the northeasterly direction, which is consistent with the permeability trends already determined from geological studies. Figure 4.49 is one of a series of maps generated from

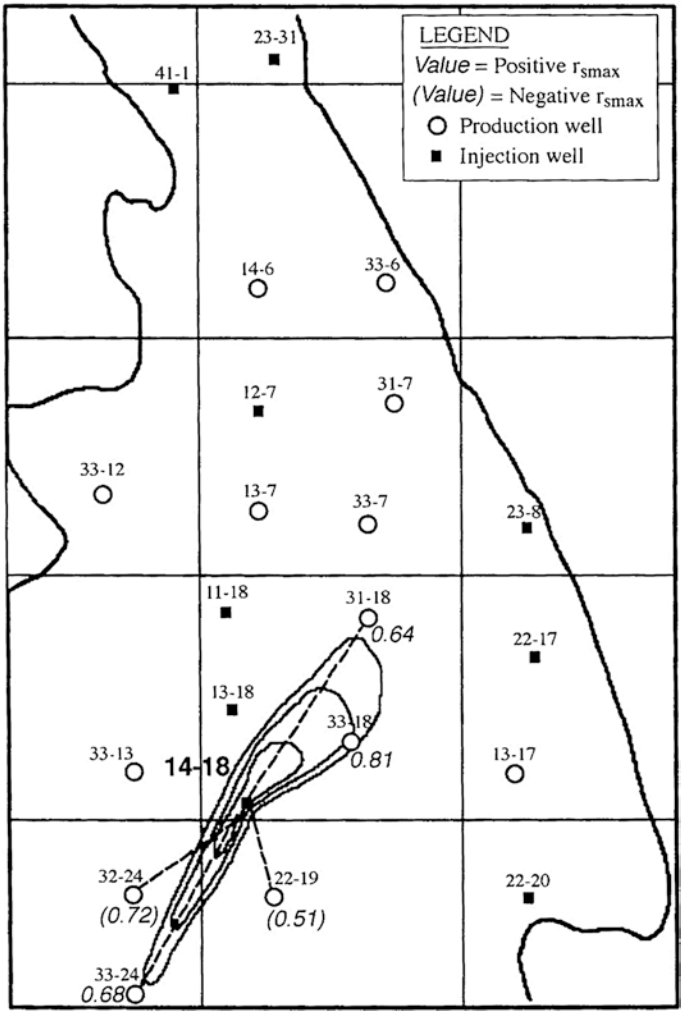


Fig. 4.49 Map of correlation values between injector and producer, plotted assuming perfect correlation (a value of 1) if the injector and producer were coincident (Refunjol and Lake 1999, Fig. 10, p. 2127)

this study, in which correlation values have been plotted and contoured. Negative correlations are interpreted as the result of influence from a third well. The map shown here is characteristic of the area, in confirming the NE-SW orientation of correlation values, and hence of reservoir connectivity and original sandbody architecture.

A particularly instructive example of the information to be gleaned from an integrated data set is that illustrated in Fig. 4.50. Hardage et al. (1996) reported on a study of the Frio gas trend in the Stratton field, Texas. A seiscrop section suggested the presence of several intersecting or overlapping channels. A suite of wells drilled through this succession confirmed the presence of channel sands, but it was impossible to interpret which, if any, are interconnected. The key data set was provided by the pressure-depth data. Pressure decline curves (Fig. 4.50b) revealed that there are three separate pressure compartments. Moreover, the almost identical decline curve for wells 127 and 161 indicated that they are in good communication. This information allowed the sand bodies to be shown as continuous between the two wells in Fig. 4.50c, and this sand body is labeled Channel C in Fig. 4.50d.

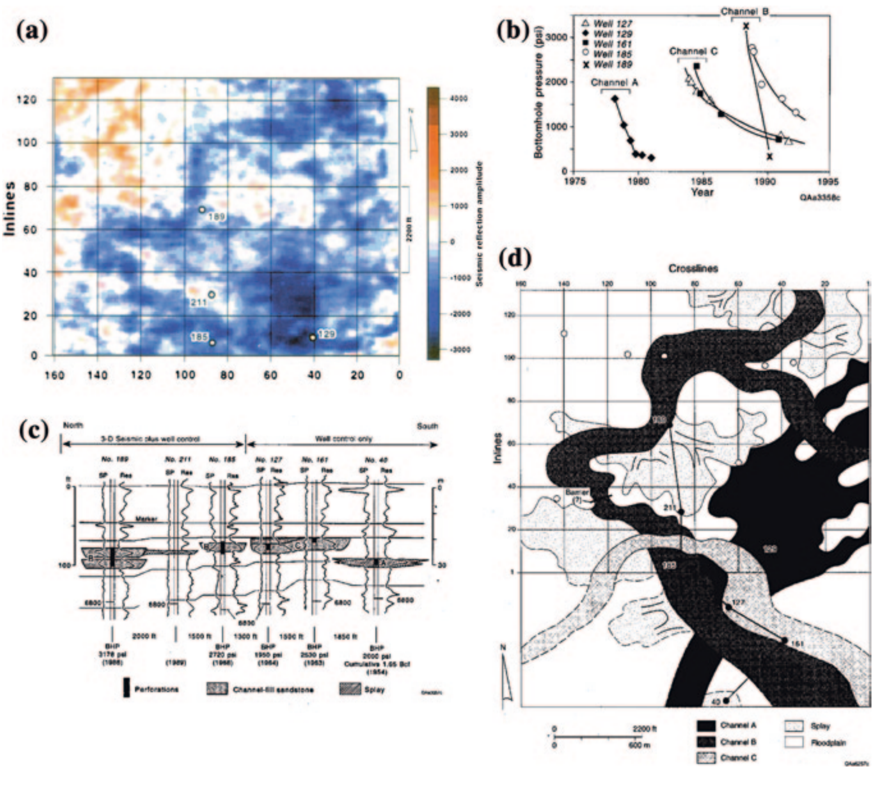


Fig. 4.50 An integrated data set. See discussion in text (Hardage et al. 1996, pp. 4, 5, 6, 7, 30, 31)

To conclude this section, survelliance methods can yield vital information regarding reservoir body connectedness, based on ongoing performance. However, such data can only be collected once a field is in production. This means that this approach cannot be used in the initial establishment of an architecture model, but may be invaluable for refining that model. Pressure data collected early in the production history (Fig. 4.50b) may become available early enough to enable the geologist to contribute sedimentological data to a reservoir model, but it may take several years before clear trends emerge. Pressure pulse tests can be carried out at any time on multiple well sets, but other trends, such as tracer tests, may also take several years to complete, given the rate at which fluids move through a reservoir.

Chapter 5
Allogenic Sedimentary Controls

5.1 Introduction

There are three major allogenic sedimentary controls: tectonism, climate change, and eustasy. All three controls may be acting on a sedimentary basin at any one time, but they are not necessarily independent (Table 5.1). Of fundamental importance is the need to recognize a distinction between upstream and downstream controls. Upstream controls include climate and tectonics. Climate controls fluvial discharge and vegetation cover and is a major influence on sediment supply. Tectonic controls affect regional slope and the relief in the source area, and therefore are a major influence in controlling the quantity and caliber of the sediment load. The major downstream control is sea-level change, except in the case of inland basins, where the downstream control is either the level of the lake into which the river drains, or the level of the tectonic rim over which the river flows out of the basin.

Shanley and McCabe (1994, Fig. 6) presented a conceptual diagram (adapted here as the upper part of Fig. 5.1) in which they illustrated, qualitatively, how allogenic controls of river systems change from river mouth to source. Clearly, at or close to sea level, base level change is the most important control. Near the source area, upstream controls will predominate. There has been much study and debate regarding the extent to which sea-level control extends upstream from a river mouth. This is discussed further, below.

Holbrook et al. (2006) introduced the useful concepts of *buttresses* and *buffers* to account for longitudinal changes in fluvial facies and architecture upstream from a coastline. A *buttress* is some fixed point that constitutes the downstream control on a fluvial graded profile (Fig. 5.1). In marine basins this will be marine base level (sea level). In inland basins it will be lake level, or the lip or edge of a basin through which the trunk river flows out of the basin. The *buffer zone* represents the available (potential) *instantaneous preservation space* for the fluvial system (Fig. 5.1). The lower limit is set by the maximum depth of local channel scour, and the upper buffer limit is the height to which the river can aggrade under

A. Miall, *Fluvial Depositional Systems*, Springer Geology,
DOI: 10.1007/978-3-319-00666-6_5,

Table 5.1 Allogenic processes affecting nonmarine environments

Process	Time scale	Effects on nonmarine processes
Plate-tectonic migration of continental plates	10^{7-8} years	Climate changes caused by changes in latitudinal position and by changes in atmospheric and oceanic circulation. Climate affects temperature, fluvial discharge, sediment yield
Plate-tectonic extension, convergence and collision	10^{7-8} years	Vertical movement of continents affects paleogeographic slopes, source-to-basin relief, and hence erosion rates and sediment delivery
Changing global average rate of sea-floor spreading	10^{7-8} years	The major cause of long-term eustatic sea-level change
Regional tectonic processes (rift faulting, nappe emplacement, etc.)	10^{4-7} years	Changes in elevation of source areas and source-to-basin relief. Generation of clastic wedges: "tectonic cyclothems"
Orbital forcing	10^{4-5} years	Climate changes, eustatic sea-level changes, including glacioeustasy. Consequent effects on temperature, fluvial discharge, sediment yield

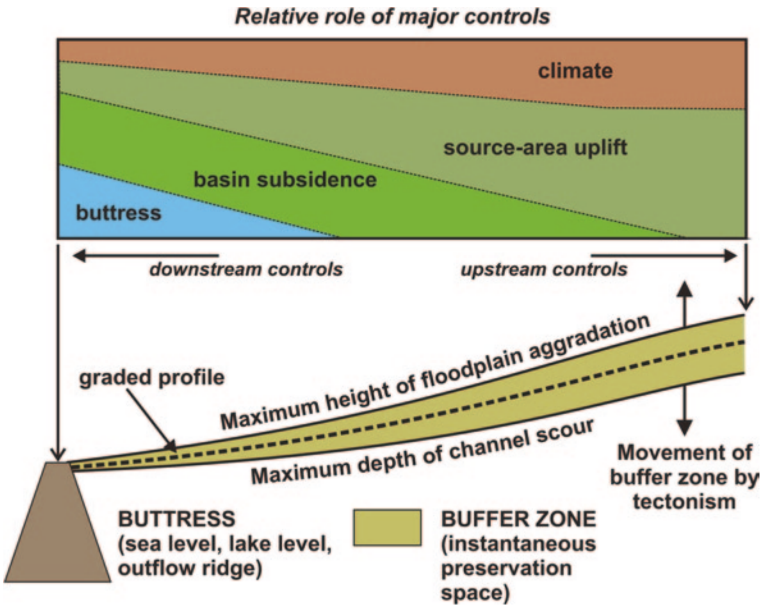


Fig. 5.1 Allogenic controls on fluvial sedimentation. The relative roles of the major depositional controls are based on Shanley and McCabe (1994, Fig. 6); the diagram is intended to suggest how the balance between upstream (tectonic, climatic) and downstream (base-level) controls changes from river mouth to source. The buttress and buffer concepts are based on Holbrook et al. (2006), and are discussed in the text

the prevailing conditions of discharge and sediment load. Scour depths can be considerable. Best and Ashworth (1997), based on their studies of the Jamuna River in Bangladesh, suggested that it may be as great as five times the mean channel depth. The buffer zone may move vertically (up or down) or expand or contract in reaction to changes in upstream controls, such as tectonism or climate change that govern the discharge and sediment load of the river. For example, tectonic uplift may increase the sediment load, causing the river to aggrade towards its upper buffer limit. The limit itself may move upward. A drop in the buttress, as a result of a fall in sea (or lake) level, may result in incision of the river system, but if the continental shelf newly exposed by the fall in sea level has a similar slope to that of the river profile, there may be little change in the fluvial style of the river. In any of these cases, the response of the river system is to erode or aggrade towards a new dynamically maintained equilibrium profile that balances the water and sediment flux and the rate of change in *accommodation*.

Away from the tectonically active basin margin, the interpretation of fluvial depositional systems in terms of tectonic controls involves a focus on the configuration, modification and rate of change of the *buffer*. This can be approached through detailed stratigraphic studies of the architecture of the deposits, including their facies variability, cyclicity, sequence stratigraphy, and changing sedimentation rates. Allen (2008, p. 20), concluded from a study of landscape evolution that

> Large alluvial systems with extensive floodplains should therefore strongly buffer any variations in sediment supply with frequencies of less than 10^{5-6} years. This has strong implications for the detection of high-frequency driving mechanisms in the stratigraphy of sedimentary basins.

It has long been known that scales and rates of clastic sedimentation vary over many magnitudes of time scale and physical scale (e.g., Sadler 1981). Recent research has indicated that this variability conforms to a fractal pattern, and this provides a useful theoretical framework from which to develop a deeper understanding of stratigraphic processes (Sect. 2.1; Miall in press). What follows are some examples of how a focus on the key features of fluvial depositional systems can contribute to these advances. Reference is made to the *Sedimentation Rate Scales* (*SRS*) defined in Sect. 2.1, which are provided in order to position discussions of sedimentation rate within a time scale appropriate for the processes under consideration.

5.2 Upstream Controls

5.2.1 Tectonics

The importance of tectonism, as a major upstream control, has long been recognized. Pettijohn (1957), King (1959) and Sloss (1962) used the term *clastic wedge* to refer to the thick, syntectonic, wedge-shaped deposits derived from orogenic uplift, and the term has subsequently been widely used for this sedimentary association. It is common for proximal deposits to make up large-scale coarsening-upward cycles tens to hundreds of meters thick, recording increasing source-area relief and depositional slope during tectonism. These have been called *tectonic cyclothems* (Blair and Bilodeau 1988; see Fig. 5.2). A direct association between tectonism and clastic wedge progradation has long been assumed but, in the case of foreland basins, was challenged by Heller et al. (1988).

Embry (1990, p. 497) proposed a set of guidelines for identifying tectonism as a major control of sequence generation. His proposal was based on his examination of the Mesozoic stratigraphic record of Sverdrup Basin, Arctic Canada, a succession up to 9 km thick, which he subdivided into 30 stratigraphic sequences in the million-year frequency range. He suggested, however, that the guidelines have general applicability. They are as follows:

> (1) the sediment source area often varies greatly from one sequence to the next; (2) the sedimentary regime of the basin commonly changed drastically and abruptly across a sequence boundary; (3) faults terminate at sequence boundaries; (4) significant changes in subsidence and uplift patterns within the basin occurred across sequence boundaries; and (5) there were significant differences in the magnitude and the extent of some of the subaerial unconformities recognized on the slowly subsiding margins of the Sverdrup Basin and time equivalent ones recognized by Vail et al. (1977, 1984) in areas of high subsidence.

To these points could be added: (6) sequence architecture can be genetically related to the developments of structures within a basin; (7) truncation of entire sequences beneath sequence boundaries indicates tectonic influence (e.g., Yoshida et al. 1996); and (8) a sequence architecture consisting of thick clastic wedges cannot be generated by passive sea-level changes (Galloway 1989).

The many types of tectonic influence on fluvial depositional processes were discussed by Miall (1996, Chap. 11), including a discussion of the Heller et al.

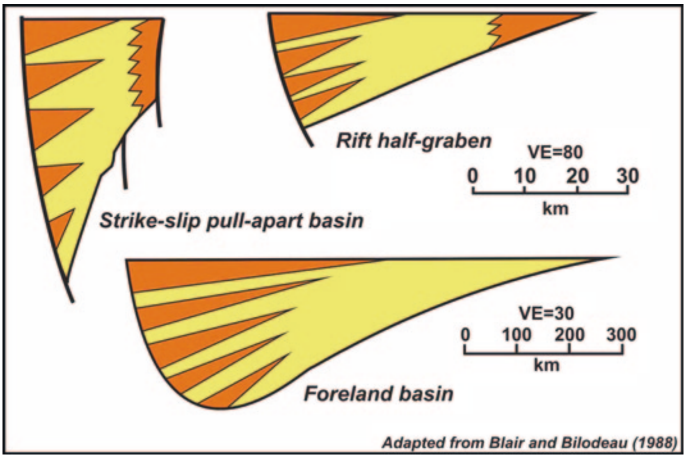


Fig. 5.2 Tectonic cyclothems. Adapted from Blair and Bilodeau (1988)

(1988) models of *syntectonic* and *antitectonic* sedimentation (Sect. 11.4.6). Schumm et al. (2000) also provided a detailed and well-illustrated dissertation on the relationships between *Active tectonics and alluvial rivers* (the title of the book). Burbank et al. (1996) documented the close relationship between active tectonism and the evolving geomorphology of river systems. Modern basin research, using the full range of chronostratigraphic dating and correlation methods can now lead to detailed interpretations of time scales and rates of processes. This section is concerned primarily with the response of fluvial systems to tectonic forcing on a regional scale. The reader is referred to the references cited above for detailed discussions of the local syndepositional response to basin-margin and within-basin tectonism.

The concept of the clastic wedge has been a very useful one, in relating large-scale nonmarine sedimentary processes to regional uplift. Other geological scenarios that contribute to an upstream tectonic control of sedimentation include syndepositional movement on bounding or within-basin faults and folds.

Clastic wedges typically approximate a wedge shape because they splay out from a point source of limited areal extent (reflecting the localization of most tectonic episodes) and thin down dip as transport energy diminishes. The term clastic wedge has come to be more broadly applied to other tectonic settings, such as wedges of alluvial-fan deposits banked against a bounding fault in a rift basin. The original concept included the supposition that the geometry and timing of clastic deposits derived from the erosion of orogenic uplifts can be correlated to tectonic episodes in their source area (the term tectonic cyclothem highlights this concept). For example, alluvial to shallow-marine Cenozoic clastic wedges of the Gulf Coast can be correlated with tectonism in the headwaters regions of the Gulf Coast rivers—the Cordilleran mountains of the western United States, with uplift and erosion along the Appalachian mountains, and with late Cenozoic glacial sediment supply from a northern hinterland (Canada).

Galloway (1989, 2005) compiled data on the ages of tectonic episodes in these source areas, and on the distribution of remnant cratonic clastic sheets in the US interior that represent areas of temporary sediment storage on the course of rivers that ultimately drained into the Gulf of Mexico. He identified nineteen genetic depositional sequences in the Gulf spanning the Paleocene to Pleistocene, and eight long-lived, extrabasinal fluvial-deltaic axes through which the bulk of the clastic detritus was delivered to the Gulf basin. From the Paleocene to the Oligocene, the bulk of the detritus was derived from Cordilleran uplifts (Fig. 5.3). The Ogallala Formation (Miocene), a fluvial sheet that extends from Wyoming and South Dakota to Texas (and serves as a vital aquifer in this relatively arid area of the High Plains) represents part of the sediment derived from the rising Cordillera and deposited in easterly and southeasterly flowing rivers that drained out into the western part of the Gulf basin. In the Miocene, rejuvenation of the Appalachian Mountains contributed to the establishment of the Mississippi drainage as a principal route for sediment delivery to the Gulf. “A broad belt of coarse, gravelly sand, derived from the southern Appalachians now forms a broad, dissected and poorly dated alluvial apron” along the southern Appalachians (Galloway 2005,

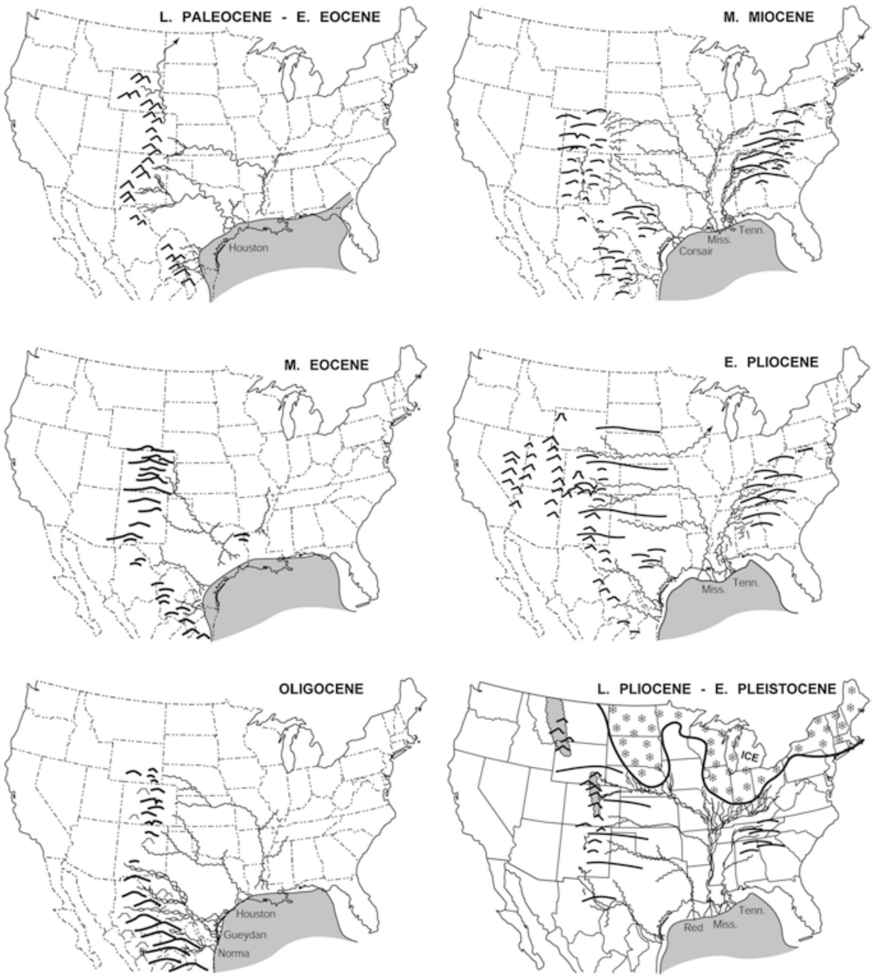


Fig. 5.3 The evolution of fluvial systems in the interior of the United States in response to orogenic and epeirogenic uplifts and (in the Plio-Pleistocene) glaciation in the north (Galloway 2005, Fig. 6, p. 418)

p. 420). The Citronelle Formation, a gravelly sand veneer along the east-central Gulf coastal plain, represents sediment derived from the Appalachians during the Pliocene, much of which entered the Mississippi or Tennessee drainage and was ultimately delivered to the Gulf. Later in the Pliocene, the flood of glaciofluvial drainage from Canada substantially increased the discharge and sediment load of the Mississippi system. Deepening and incision of this river led to its capture of the Red and Tennessee rivers in the Pleistocene.

Another example of the correlations that can be made between tectonism and sedimentation encompasses the various pulses of nonmarine sediment that characterize the interior basins of British Columba and the clastic wedges that entered the Alberta foreland basin from the rising Cordilleran uplifts during the Jurassic-early Tertiary (e.g., Kootenay-Fernie, Blairmore, and Belly River-Paskapoo: see Fig. 5.4). The ages of these local basin fills and foreland basin clastic pulses have been loosely correlated with terrane-accretion events along the western continental margin (Stockmal et al. 1992; Ricketts 2008). Petrographic and paleocurrent evidence has been crucial in these reconstructions. Timing of terrane docking has, in many cases, been confirmed by the presence of "overlap assemblages" of detritus that was clearly derived from one terrane (based on its detrital petrography), overlaps the terrane suture and extends onto the adjacent terrane.

Clastic wedges and tectonic cyclothems occur over a wide range of physical scales and time scales. For example, Figs. 5.5, 5.6 and 5.7 illustrate part of the Sevier clastic wedge of the Rocky Mountain foreland basin at successively larger scales. The entire wedge (Fig. 5.5) represents a 30 m.y. (Cenomanian to Maastrichtian) cycle of progradation some 500 km across, from the Wasatch Range of central Utah, to the Front ranges near Denver. Figure 5.6 illustrates the lower part of this wedge, consisting of three major cycles spanning approximately 10 m.y. The Ferron Sandstone, deposited during the mid- to late Turonian, representing perhaps 3 m.y., constitutes the most basinward extent of coastal regression during this period. In detail (Fig. 5.7) it can be seen that the Ferron comprises nine cycles of coastal regression, each followed by transgression and coal accumulation. Ryer (1984) attributed the higher order cyclicity shown in Figs. 5.6 and 5.7 to eustatic sea-level change, based on purported correlations with the global cycle charts of Vail et al. (1977). However, studies of part of the Upper Cretaceous record of the Rocky Mountains basin indicate that eustatic sea-level control is unlikely, based on the lack of correlation between cycles at widely spaced sections within the basin (Krystinik and DeJarnett 1995). Miall and Miall (2001) and Miall (2010) suggested that the Vail/Exxon curves should not be accepted as standard scales for the determination of eustatic control. It is more likely that that the million-year cycles illustrated in Fig. 5.6 are tectonic in origin. The presence of an unconformity landward of the Ferron sandstone tongue is consistent with an interpretation that the Ferron sandstone is the product of tectonic rejuvenation of the source area. The high-frequency Ferron cycles of Fig. 5.7 may be the product of orbital climatic or sea-level forcing, a mechanism discussed below.

Some of the most detailed work relating sedimentation to active tectonism has been carried out in foreland basins, particularly the Sevier clastic wedge of the Rocky Mountain Basin and the basins flanking the Pyrenees. The physical scale of these cycles, and the geologic time that they represent, are reflections of the processes involved in their generation. For example, Table 5.2 summarizes the local to regional tectonic processes that may occur in foreland basins, ranging from the movement and loading of individual thrust plates to the crustal loading caused by the collision and suture of an entire continental margin.

Growth strata that develop adjacent to active structures, such as basin-margin thrust faults are typically deposited at *SRS-10* rates, which represent the highest long-term geological rates that have been recorded. Along the southern flanks of

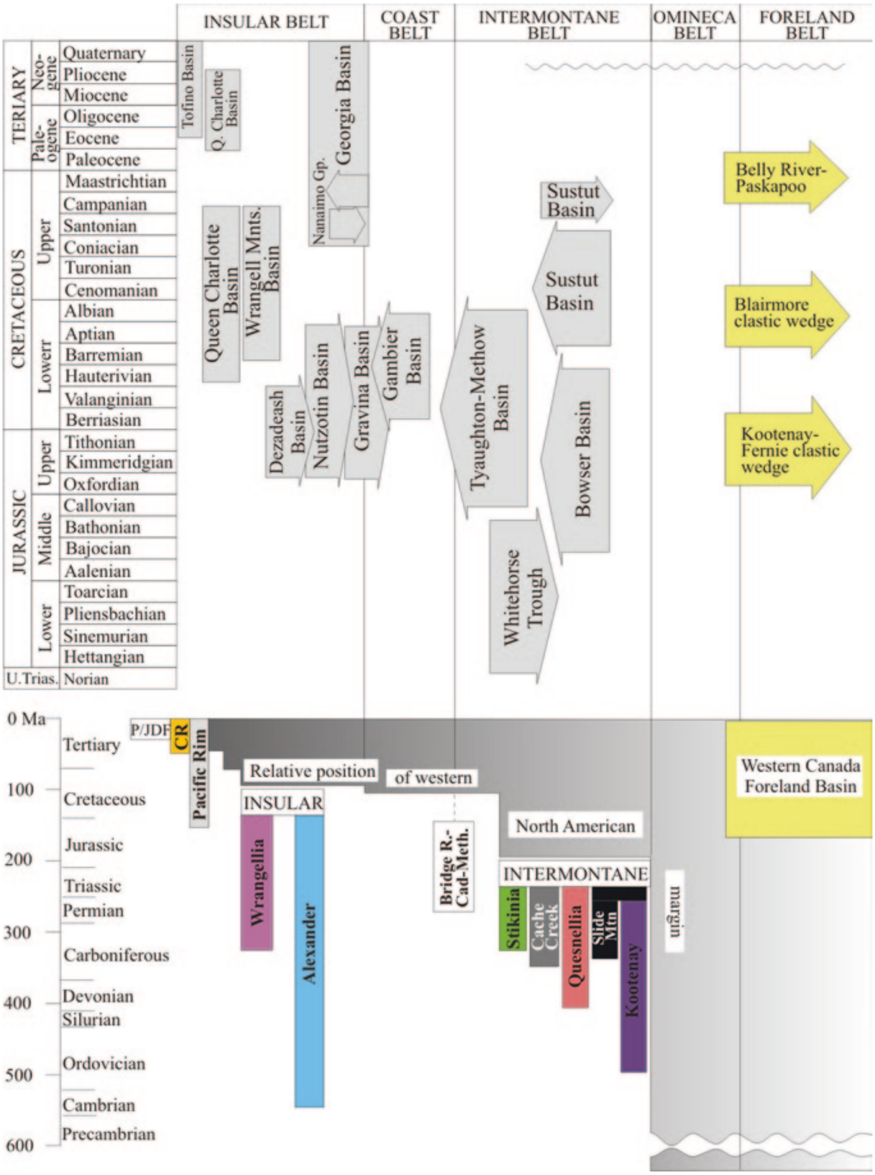


Fig. 5.4 *Top*: Space–time relationships amongst the major sedimentary basins of the Canadian Cordillera. Large *arrows* show the generalized direction of sediment flux, reflecting relative uplift of source areas to the west or east. The three clastic wedges indicated at right (*yellow arrows*) are the major clastic pulses that entered the foreland basin in Alberta. *Bottom*: *Coloured rectangles* indicate ages of the major terranes that comprise the Rocky Mountains of British Columbia. The times of amalgamation of the terranes into “superterranes” are indicated by the position of the names of the two superterranes, Insular and Intermontane (Ricketts 2008, Fig. 3, p. 369)

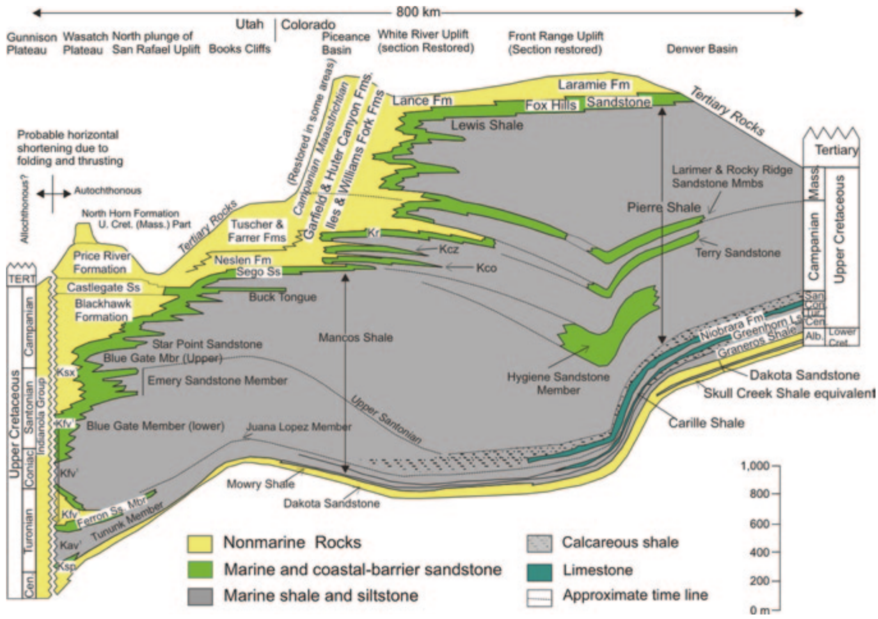


Fig. 5.5 The Sevier clastic wedge, Utah-Colorado. Adapted from Molenaar and Rice (1988)

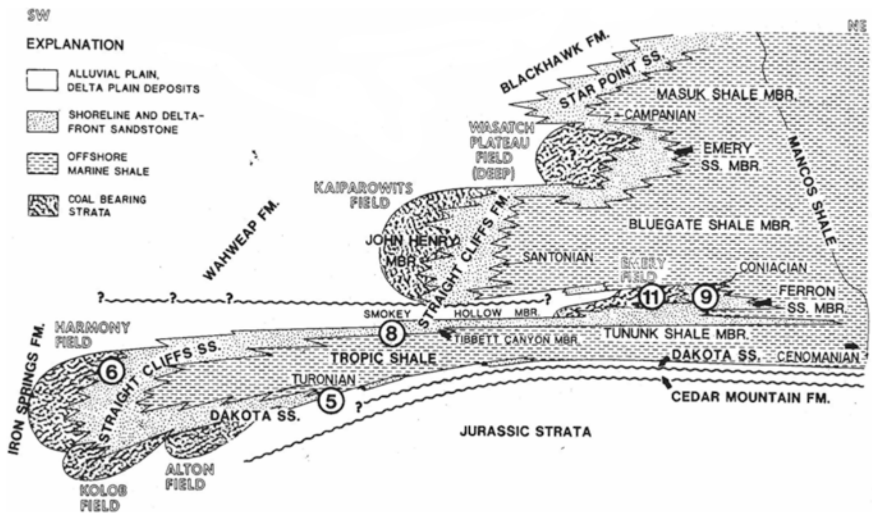


Fig. 5.6 The lower part of the Sevier clastic wedge (Ryer 1984, Fig. 1, p. 218)

the Pyrenees coarse, proximal, syntectonic alluvial deposits are well exposed and detailed mapping has revealed the patterns of uplift, erosion, basement unroofing, and coarse clastic sedimentation (Barrier et al. 2010). Unfortunately, it is very difficult to accurately determine the rates of processes in this type of setting, owing

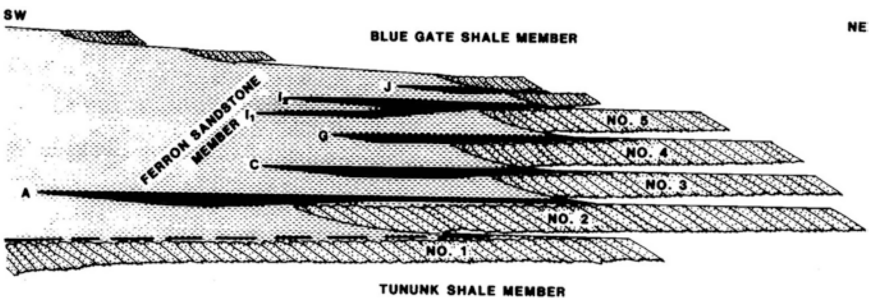


Fig. 5.7 The minor cycles of the Ferron Sandstone (Ryer 1984, Fig. 4, p. 220)

Table 5.2 The relationship between tectonic processes and stratigraphic signatures in foreland basins, at different time scales

Duration m.y.	Scale	Tectonic process	Stratigraphic signature
>50	Entire tectonic belt	Regional flexural loading, imbricate stacking	Regional foredeep basin
10–50	Regional	Terrane docking and accretion	Multiple "molasse" pulses
10–50	Regional	Effects of basement heterogeneities during crustal shortening	Local variations in subsidence rate; may lead to local transgressions/regressions
>5	Regional	Fault-propagation anticline and foreland syncline	Sub-basin filled by sequence sets bounded by major enhanced unconformities
5–0.5	Local	Thrust overstep branches developing inside fault-propagation anticline	Enhanced sequence boundaries; structural truncation and rotation; decreasing upward dips; sharp onlaps thick lowstands, syntectonic facies
<0.5	Local	Movement of individual thrust plates, normal listric faults, minor folds	Depositional systems and bedsets geometrically controlled by tectonism and bounded by unconformable bedding-plane surfaces. Maximum flooding surfaces superimposed on growth-fault scarps. Shelf-perched lowstand deposits

This table was adapted mainly from Deramond et al. (1993), with additional data from Waschbusch and Royden (1992), Stockmal et al. (1992)

to the difficulty in determining accurate ages of coarse alluvial deposits. Burbank et al. (1996, Fig. 6) provided an example where accumulation rates averaged 0.117 m/ka over 1.7 m.y. Data provided by Medwedeff (1989) indicate a growth rate of 0.305 m/ka over 8 m.y. for an example in California. At the margins of the Tarim Basin, in western China, Sun et al. (2010) used magnetostratigraphic data to determine the rates of accumulation of "growth strata" in proximity to a growing anticline. Sedimentation rates increased from 0.325 m/ka prior to the syndepositional movement of the growth structure, to 0.403 m/ka during the period of active tectonism, over a total time span of about 10 m.y.

In some cases, tectonic episodes can be linked quite specifically to depositional episodes. Mapping by R. L. Armstrong, P. G. DeCelles, and many others since the 1960s has unraveled a close relationship between episodes of thrust faulting and uplift along the Sevier fold-thrust belt in the US Rocky Mountains region, and the development of coarse, source-proximal conglomerates derived directly from those uplifts (e.g., Horton et al. 2004). Eastward, the conglomerates pass into nonmarine and shallow-marine coastal plain deposits. Commonly the conglomerates are themselves cut and displaced by the faults, which is further evidence of the genetic relationship between tectonism and sedimentation. Figure 5.8 summarizes this relationship, as it has been documented in northeastern Utah and southern

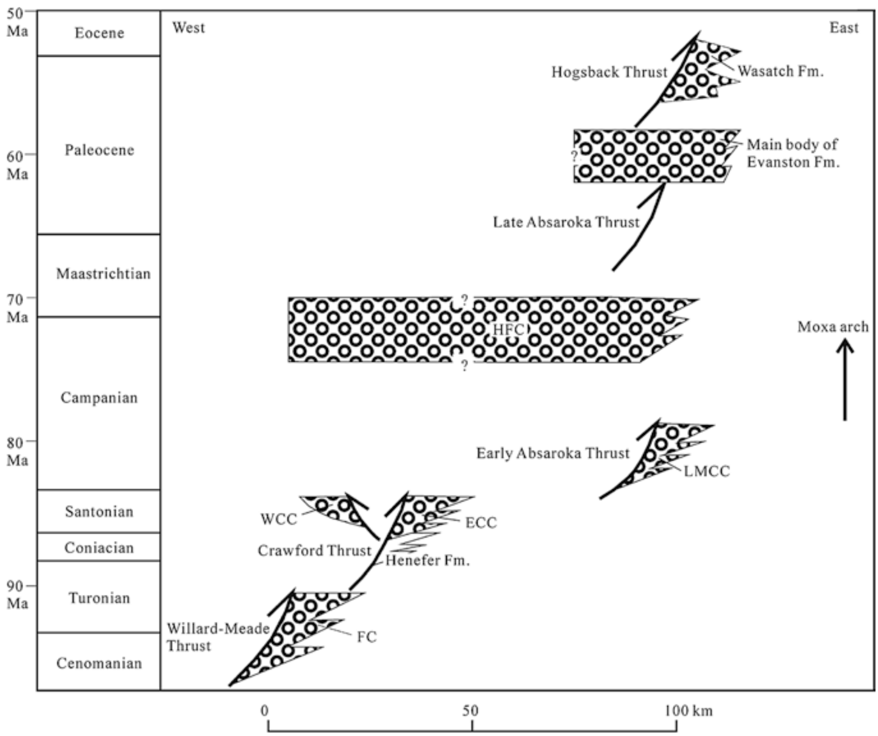


Fig. 5.8 The time–space relationship between thrust fault episodes and synorogenic conglomerate deposition in northeastern Utah and southern Wyoming. *HFC* Hams Fork Conglomerate; *LMCC* LittleMuddy Creek Conglomerate;*WCC* Weber Canyon Conglomerate; *ECC* Echo Canyon Conglomerate; *FC* Frontier Conglomerate (Liu et al. 2005, Fig. 3, p. 491)

Wyoming, and Fig. 5.9 is a synthesis of the stratigraphy of these sequences. Figure 5.10 is a chronostratigraphic chart which synthesizes the stratigraphy, facies and timing of these sequences in relationship to regional and global events.

The definition of the sequences shown in Figs. 5.9 and 5.10 did not automatically emerge from a synthesis of the regional stratigraphy, but required a search for and a recognition of the key indicators of changing regional accommodation that are now know to characterize sequence generation. This is a good illustration of the "genetic" nature of sequence stratigraphy—application of sequence concepts can greatly facilitate synthesis and interpretation, but only if there are existing concepts and models that are appropriate for the field case under study. This illustrates both the strengths and the pitfalls of the method. In the case under study here, sequence mapping was facilitated by the appropriate choice of datum for the construction of the stratigraphic synthesis (Fig. 5.9). This is not a mundane methodological issue, but may become a key to the elucidation of stratigraphic relationships. The datum, in this case, was placed at the base of the Canyon Creek Member of the Ericson Formation, which clarifies the progradational nature of the units above, and introduces as little distortion as possible to the complex tectono-stratigraphic relationships of the units below. The five sequences into which the succession has been divided were recognized in the basis of "an iterative process,

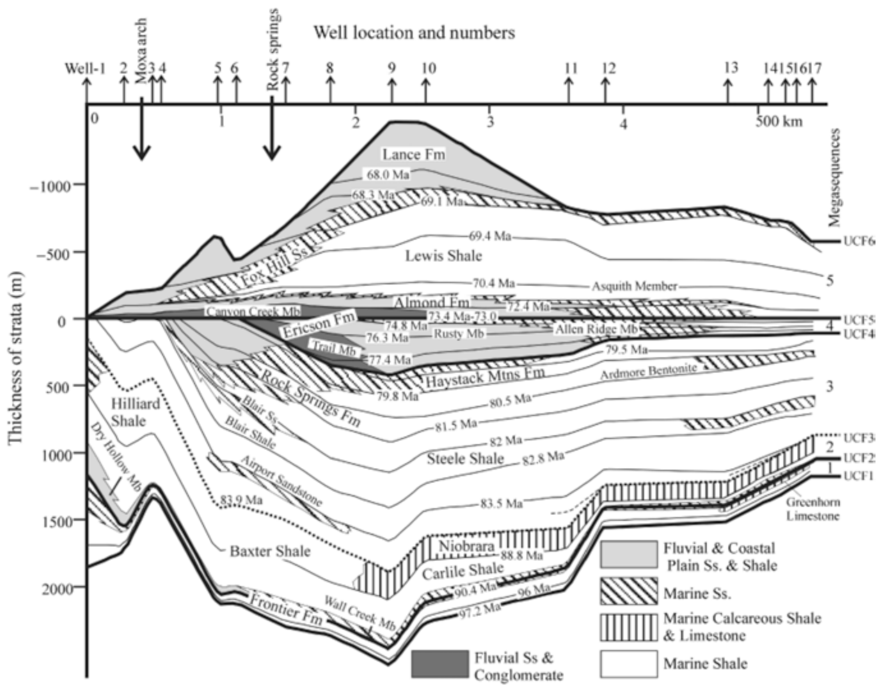


Fig. 5.9 The stratigraphy of Upper Cretaceous (mid-Cenomanian-Maastrichtian) "megasequences" in southern Wyoming. *UCF* = unconformity (Liu et al. 2005, Fig. 4, p. 493)

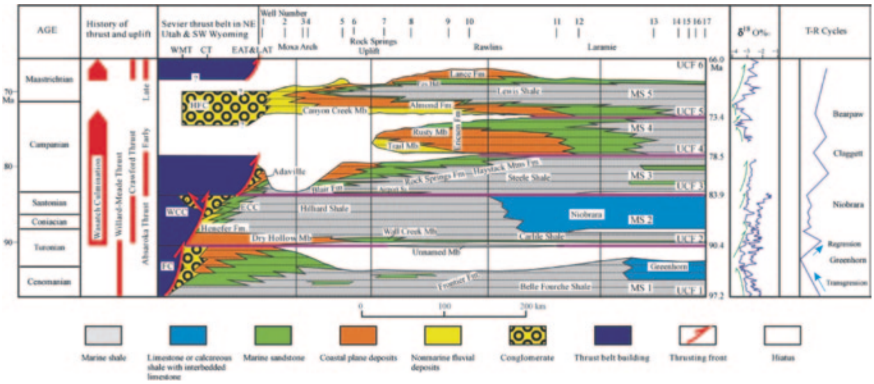


Fig. 5.10 Correlation chart for the Upper Cretaceous "megasequences" of southern Wyoming, showing facies variations, relationship to thrusting episodes and, at right, the oxygen isotope curve from Abreu et al. (1998) and the T-R cycles of Kauffman (1984). *WMT* Willard-Meade thrust; *CT* Crawford thrust; *EAT* Early Absaroka thrust; *LAT* Late Absaroka thrust. *MS* = megasequence, *UCF* = unconformity (Liu et al. 2005, Fig. 5, p. 494)

in which regional unconformities and/or surfaces across which there was a demonstrable rapid change in subsidence regime were the chosen boundaries" (Liu et al. 2005, p. 493). Correlation of the units westward, through the facies change into the syntectonic conglomerates, was also an important criterion. Using accommodation cycles as a model for interpretation, lithostratigraphic units that have long been known in this area could be assigned their appropriate position in the succession of systems tracts. Specific marine shales could then be identified as transgressive deposits or representing maximum flooding intervals, upward-coarsening transitions from coastal-plain sandstone to coarse conglomerate could be assigned to highstand systems tracts, and so on. The complete Cenomanian-Maastrichtian succession took 32 m.y. to accumulate to a maximum thickness of 3,500 m (Fig. 5.9), a sedimentation rate of 0.108 m/ka, a rate consistent with *SRS 10 or 11*.

Aschoff and Steel (2011) summarized earlier work on the Sevier clastic wedge, in the classic area of the Book Cliffs (Utah-Colorado), focusing on the Campanian section (Fig. 5.11; the middle portion of the Sevier wedge, as illustrated in Fig. 5.5). This succession has long been interpreted as the product of repeated thrust loading along the Sevier fold-thrust belt (Kamola and Huntoon 1995; Yoshida et al. 1996; Horton et al. 2004). Aschoff and Steel (2011) calculated rates of coastal progradation and rates of sediment accumulation in order to explore relationships between sedimentation and tectonism. The range of sedimentation rates is 0.047–0.14 m/ka, calculated over stratigraphic times spans of between 2.1 and 6.5 m.y. These are within the range for *SRS 11*, but are low relative to those recorded in some Andean basins (Miall, in press). By comparison, the Catskill Delta of New York-Pennsylvania accumulated at comparable rates. Data provided by Ettensohn (2008) indicate a maximum rate for the proximal part of the "delta" (in reality a major clastic wedge deposited in a range of nonmarine to

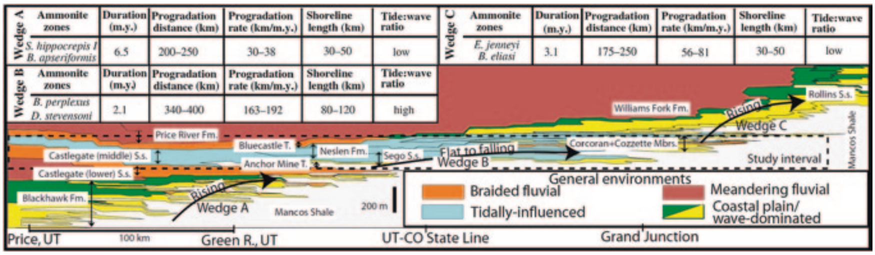


Fig. 5.11 Summary of Campanian clastic wedges in the Cordilleran foreland basin showing units, shoreline stacking trajectories, generalized facies types, and progradation rates (Aschoff and Steel 2011, Fig. 2, p. 1825)

shallow-marine environments) of 0.096 m/ka (maximum thickness of 3 km accumulated over about 9 my between the Givetian and the Famennian).

Horton et al. (2004) cited petrographic evidence for tying the origin of the Castlegate Sandstone to movement on thrust structures in the Sevier orogen. Aschoff and Steel (2011) speculated about the possible influence of basement uplift within the Sevier foreland basin, which would tend to cancel out some of the subsidence due to flexural loading. This seems particularly likely for the middle portion of the clastic wedge, that characterized by the Castlegate sandstone, the sheet-like nature of which has, for some time, been attributed to a slow rate of regional subsidence (Yoshida et al. 1996). The incipient activation of Laramide structures within the basin, as suggested by Aschoff and Steel (2011), would be consistent with these characteristics of the clastic wedge. However, no such special influence on rates of accommodation has been suggested for the Appalachian basin.

Distinguishing downstream eustatic from upstream tectonic causes of cyclicity in the fluvial record depends on the construction of a regional overview, with correlation to tectonic or eustatic episodes in the upstream or downstream directions, respectively. Two additional examples are described here to highlight the stratigraphic and sedimentologic features that can cement the relationship between sedimentation and tectonics.

Herrero et al. (2010) described the evolution of a largely subsurface foreland basin of Cenozoic age in northern Spain. Well records (Fig. 5.12) display characteristic conglomeratic-sandy autogenic cycles. Seismic reflection data reveal the presence of three major unconformity-bounded sequences, in which the angularity of the unconformity and the depth of erosion increase toward the source area (Figs. 5.13 and 5.14). Paleocurrent data confirm the uplifted basin margin as the sediment source, to the north of the basin. Thickness and age information for the Vegaquemada sequence, the oldest of the three, indicate that it was deposited over a period of 25 m.y. and reached a maximum thickness of 1,400 m in the north, immediately adjacent to the main bounding fault. This translates into a long-term accumulation rate of 0.056 m/ka, characteristic of *SRS-11*, the rate typical for basin-fill complexes undergoing average regional subsidence rates.

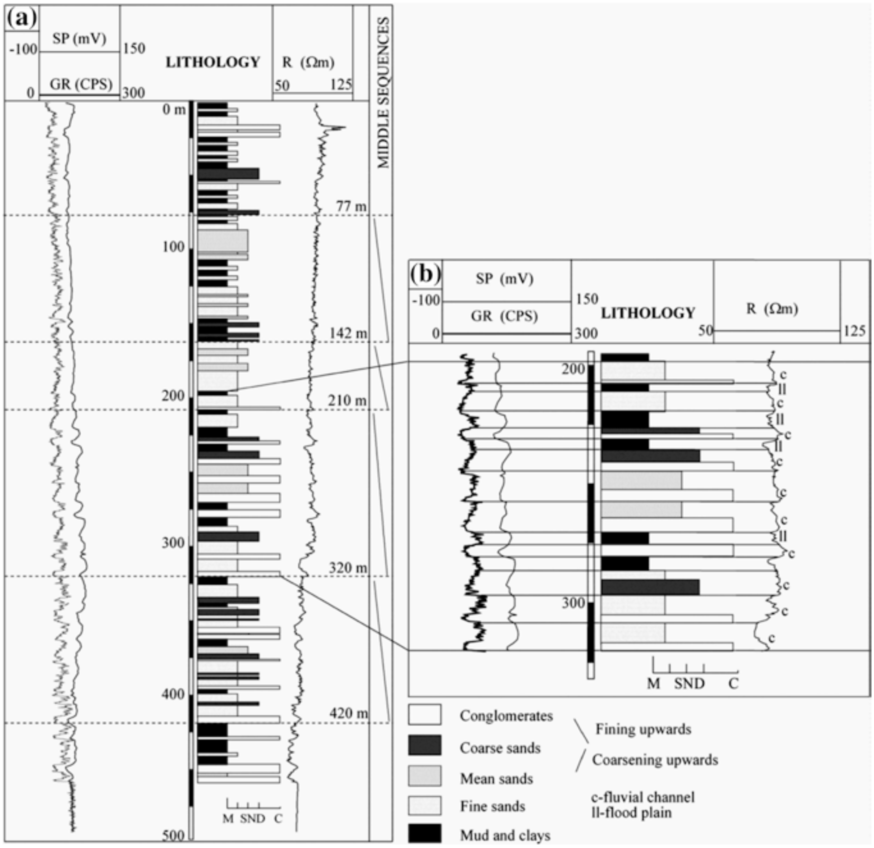


Fig. 5.12 Well log through part of the Candanedo sequence (Miocene), Duero basin, northern Spain, showing a succession of autogenic fluvial cycles (Herrero et al. 2010, Fig. 13, p. 255)

Turning to a completely different tectonic environment, Mack et al. (2006) documented in detail the sedimentary history of the southern Rio Grande Rift in New Mexico. During a largely aggradational phase, extending from 5 to 0.8 Ma, approximately 100 m of fluvial sediment were deposited in eight separate sub-basins along the rift system (sedimentation rate 0.023 m/ka, *SRS 9*). Most of this sediment accumulated along the central rift, where it was deposited by the axial ancestral Rio Grande River (Fig. 5.15). Alluvial fans prograding from the faulted basin margins deposited narrow wedges of coarse conglomerate. The variation in stratigraphic completeness from basin to basin, the presence of internal disconformities and mature paleosoils attest to the movement on individual faults within the rift system, causing local differential uplift and subsidence movements. Leeder (e.g., Leeder and Gawthorpe 1987, and later papers) has documented in detail the relationships between sedimentation and tectonics in rift systems characterized by subsidiary and cross-cutting faults, how the tectonic evolution of these structures

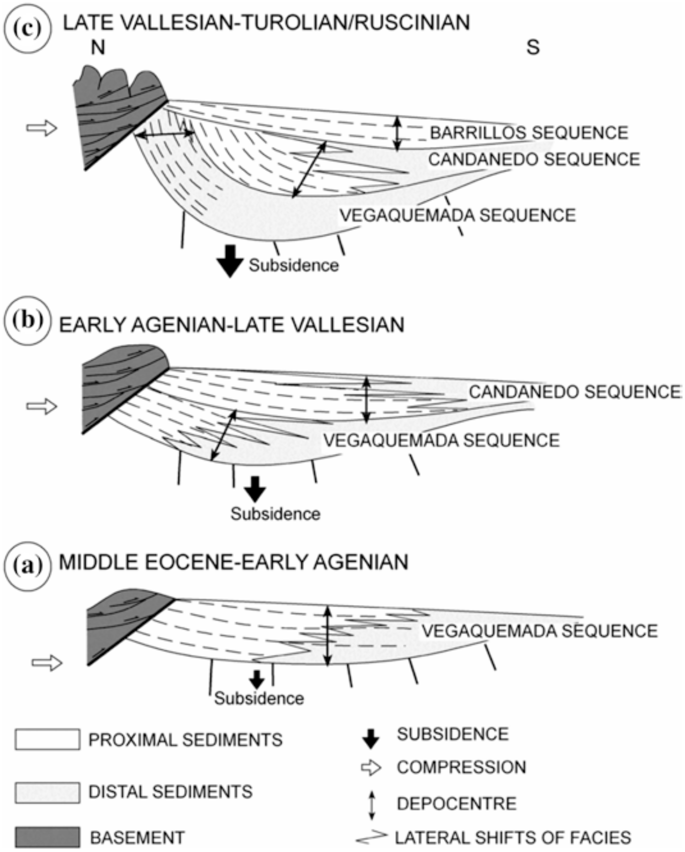


Fig. 5.13 Evolution of the three major tectono-stratigraphic sequences in the Duero basin, Spain (Herrero et al. 2010, Fig. 19, p. 260)

controls the development of steep erosional scarps on the face of downfaulted blocks, and the location of marginal erosional valleys and sediment delivery systems.

López-Gómez et al. (2010) described the fluvial fill of a Permian–Triassic basin in eastern Spain that subsided at varying rates as a result of episodic extensional faulting and thermal subsidence. The most rapid subsidence rates, in the range of 25 m/yr, occurred over a 2 m.y. period in the late Permian (256–254 Ma; López-Gómez et al. 2010, Fig. 6). Subsidence rates during the Triassic were at about half this rate, but still, at 10^{-2} m/ka, within the *SRS-11* range. This study explored the relationship of fluvial architecture to stretching factors, using values of δ and β for upper- and lower-crustal stretching factors calculated form borehole data. The hypothesis to be tested was whether stretching factor could be related to variations between various styles of amalgamated or isolated sandbody, it being assumed that the architecture would be a proxy indicator of subsidence rates. However,

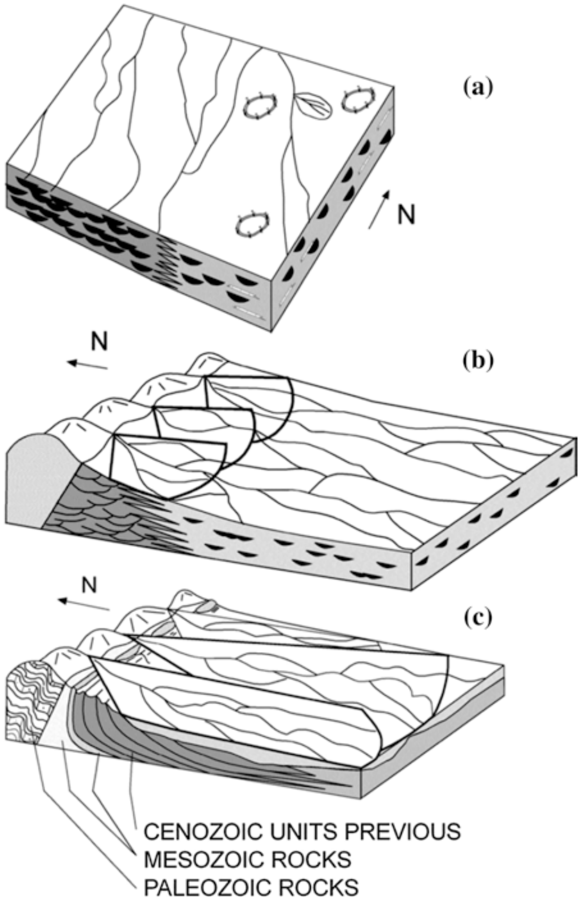


Fig. 5.14 Depositional models for the (**a**) Vegaquemada sequence, (**b**) Candanedo sequence and (**c**) Barrillos sequen, Duero Basin, Spain (Herrero et al. 2010, Fig. 21, p. 261)

subsidence rates were neither quoted nor discussed in the paper (they may be calculated from data presented in their Fig. 5.6). They concluded:

> Collectively, our field and laboratory data suggest that although general subsidence in some way controls the resultant fluvial geometry of the Permian and Triassic alluvial sediments of the Iberian Ranges, there is no simple direct relationship between the two factors. (López-Gómez et al. 2010, p. 329)

They suggested a relationship between architecture and stretching factor, but the relationships do not appear to be straightforward. Indeed, they noted that other factors, including climate change, could have affected the final composition of the preserved alluvial architecture.

It is suggested here that the relationships between subsidence and alluvial architecture are far more complex than proposed by López-Gómez et al. (2010). Alluvial architecture—the preserved complex of amalgamated macroforms, is determined by sedimentary processes that occur within the *SRS-6* to *SRS -8* range,

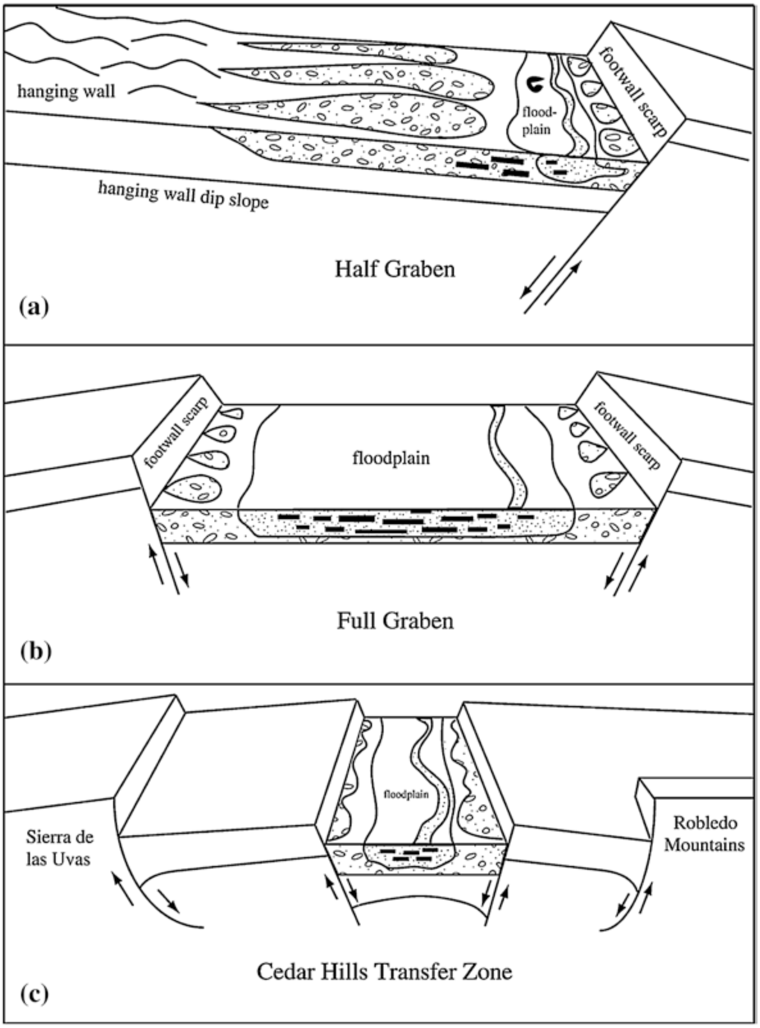


Fig. 5.15 Schematic cross-sections of: (**a**) half graben, (**b**) full graben, and (**c**) Cedar Hills Transfer Zone, showing distribution of alluvial-fan and axial-fluvial sediment of the Camp Rice and Palomas formations. *Black rectangles* in cross-sections represent floodplain and distal alluvial-fan mudstones (Mack et al. 2006, Fig. 9, p. 152)

that is, on time scales of 10^2–10^5 years and sedimentation rates of 10^{-1}–10^2 m/ka. Formation and preservation of the elements of alluvial stratigraphy are therefore completed within time scales several orders of magnitude more rapid that regional subsidence (see Sect. 6.2 and Miall, in press). This may be well illustrated by a comparison with the geology of the Rio Grande Rift, the development of which is referred to briefly above.

Research in experimental stratigraphy is providing some essential insights into the response of fluvial systems to tectonic forcing. Kim and Paola (2007) modeled the case of a fluvial system flowing over and between active extensional faults oriented normal to flow direction. When interpreted in terms of full-scale systems their experiments demonstrated that a pulse of fault movement would trigger an allogenic response that would take about 10^5 years to complete and would generate a cycle in the range of 100 m thick (*SRS 7-8*). However, autogenic adjustments to the tectonic forcing would include cycles of channel adjustment and alternations between episodes of sediment storage and release (at times of channel incision) that would generate a set of smaller-scale cycles nested within the main cycle. In their experiments, these nested cycles were typically 10–20 m thick. This is not inconsistent with the conclusions of Allen (2008), quoted in the Introduction to this chapter (Sect. 5.1). Allen (2008) summarized theoretical work and observations on modern alluvial fan-catchment systems where tectonic episodes (fault movements) and landscape response times were measured over a time scale of a few millions of years. Whereas large alluvial systems are expected to damp out tectonic forcing of sediment supply over a 10^5–10^6-year time scale, observations of small-scale river systems indicate that response times may be as short as 10^4 years.

5.2.2 Climate

The late Paleozoic Mid-Continent cyclothems were amongst the first ancient deposits to be interpreted in terms of an overall climatic control. Shepard and Wanless (1935) suggested a glacioeustatic control based on the major regional continental glaciation underway across Gondwana at the time of cyclothem deposition (Pennsylvanian-Permian). These deposits locally contain thick deltaic successions, which Wanless (1964) attributed to regional tectonism along the Appalachian orogen. The interpretation of cyclothem deposition as the result of orbitally-forced, high-frequency sea-level change associated with continental glaciation in Gondwana was a major conceptual development that has not subsequently been challenged. However, subsequent studies (most recently: Soreghan and Montanez 2008; Allen et al. 2011) have demonstrated a wide range of variability in the nature and time scale of climate change through this period.

Miall (1996, Chap. 12) discussed a range of problems that arise when attempting to identify and isolate climatic controls on sedimentation in the ancient record:

1. Large variations in fluvial discharge produce recognizable effects in the facies record, such as the interbedding of very coarse and very fine facies, but the time scale of variation may not be obvious and the climatic implications therefore unclear. Seasonal effects, such as those arising from the freeze–thaw cycle, monsoonal variations, and those that occur in arid environments characterized by occasional violent flash floods months or even years apart, all generate the same wide variation in discharge, and consequent wide variation in facies, but the climatic interpretation of the resulting facies may be quite ambiguous.

2. The climate of the sediment source area and that of the depositional basin may not be the same. For example, rivers flowing southward from the Alpine-Himalayan mountain chain are all strongly seasonal, reflecting the alpine climate of their mountainous source areas, but they enter very different climates in the plains to the south. European rivers, such as the Rhone and Po, with headwaters in the Alps, have coastal plains in the subtropical Mediterranean region. The Tigris and Euphrates enter the highly arid Mesopotamian basin. The fluvial plains of northern India are located in a tropical climate where seasonal monsoonal rainfall strongly influences discharge characteristics.
3. Related to the above point is the fact that flow hydraulics of a river, and therefore the facies and architecture of the bedload deposits, are largely determined by the climate of the source area, whereas floodplains and their finer grained deposits are influenced primarily by the climate of the depositional basin.
4. Topographic effects may generate local climates that complicate the depositional record. For example, mountain ranges rising in the path of prevailing humid winds may generate high orographic rainfalls on their upwind slopes and rain shadows downwind. The topographic diversion of air masses is an important general effect. Very high plateaus, such as Tibet, are also a significant influence on regional climatic patterns, which have their own effects on air temperatures and circulation.
5. Even within a fluvial basin, levels of water saturation (the water table) and redox conditions may vary from channel to floodplain and between the proximal, possibly more elevated basin-margin rivers and the topographically lower basin-centre channel systems.
6. Climate is commonly recorded in the sedimentary record by vegetation, the fossil remains of vegetation itself, and by the effects vegetation has on sediment erodibility, sediment yield, channel style, and so on. However, the styles of vegetation that characterize the earth today have evolved with time, and were very different in the geological past. The development of large land plants in the Devonian, of plants that could survive seasonal climate changes in the Mesozoic, and of grasses in the Miocene, all brought about changes in fluvial hydrology and channel style. This was first suggested by Schumm (1968a), and has been explored in depth and confirmed by the detailed reviews of Davies and Gibling (2010a, b, 2011). Modern analogues therefore have a limited applicability to the distant past, with pre-Devonian landscapes probably functioning sedimentologically much as do arid regions of the present day, even where rainfall was abundant.

Correctly identifying these various effects and separating them from the sometimes very similar effects of tectonic control, requires meticulous observation and very careful deductive reasoning.

The Late Paleozoic Ice Age of Gondwana, which extended from the late Pennsylvanian to the Early Permian, was a period of climate instability worldwide (Soreghan and Montanez 2008). Much of the world's coal reserves in North America and Europe were deposited during this period, at a time when these regions were situated between 30°N and 30°S, deep within the supercontinent Pangea. Partly for this reason there has been considerable interest in exploring the relationship between sedimentation and climate, and the controls on coal development. Tabor and Poulsen (2008) listed seven factors that may have influenced paleotropical climates: (1) tectonic drift, (2) land-sea distribution, (3) supercontinentality, (4) monsoon variability, (5) uplift/collapse of major orogenic belts, (6) waxing and waning of Gondwanan ice sheets, and (7) atmospheric pCO_2. How these regional and global controls affected alluvial architecture is discussed below.

The sedimentological evidence of climate change is not limited to field cases where it is possible to establish high-frequency sea-level change (of glacioeustatic origin). Cecil (1990) and Perlmutter and Matthews (1990) provided general models of the response of depositional systems to climate change. Climate controls sedimentation through its effects on temperature, humidity, rainfall, evaporation rates, wind and sunlight, all of which affect vegetation cover, weathering and erosion patterns, and sediment yield (Table 5.3; Figs. 5.16, 5.17). As climate belts shift as a result of orbital forcing, or the continents drift through climate belts as a result of plate motions, climate changes lead to changes in clastic sediment yield and patterns of chemical sedimentation. Vegetation patterns change both in elevation and latitude, as the global climate cycles through the changes in seasonality and total energy flux. Arid climates are times of low sediment yield and, in low-lying areas, are accompanied by formation of pedogenic carbonates and evaporites. Sediment yield increases with increasing precipitation, leading to a predominance of clastic sedimentation. Maximum yields occur under temperate, seasonal wet/dry climatic conditions. Very humid climates are characterized by thick vegetation cover, which results in a reduction in sediment yield and an increasing importance of peat/coal formation. Specialized studies of paleosoils can yield much information regarding prevailing climates and climate change (Retallack 2001). Leier et al. (2005) demonstrated that giant alluvial fans ("megafans") occur where monsoonal climate conditions are responsible for large-scale discharge fluctuations, resulting in the rapid dumping of enormous volumes of coarse detritus along a mountain front.

As discussed in detail by Blum and Törnqvist (2000), the sedimentological behavior of a river depends on the balance between discharge and sediment load (Fig. 5.18). Sediment supply depends on many factors, with climate being a primary influence, as noted above. Stream power is determined primarily by discharge, which may be steady or flashy, depending on climatic regime. Given a river system that has reached a dynamic equilibrium, an increased in bedload will result in aggradation, whereas an increase in discharge, or a reduction in bedload, will result in degradation.

The Niger River and delta system offers an interesting example of how changes in climate might be directly reflected in alluvial sedimentation (Fig. 5.19). The Niger River, which is the major sediment source for the delta, flows through three major climate belts that cross tropical west Africa. In the north, the Sahel is an area of steppe climate—arid, with very limited vegetation cover. Rainfall is sparse and flashy, leading to erratic yield of coarse clastic detritus. The savanna is an

Table 5.3 Responses to nonseasonal and seasonal rainfall under tropical and subtropical temperatures

Variable	Tropical rainy	Long wet/short dry	Wet-dry	Semiarid	Arid
Rainfall	High, nonseasonal	Short dry season	Extreme seasonality	Short wet season	Arid
Vegetation	Rain forest	Forests	Grasslands	Steppes	Shrubs
Chem weathering	Intense	Intense to mod	Mod to restricted	Minimal	Very low
Products, soils	High-Al clays quartz histosols latosols	Latosols histosols	Vertisols histosols?	Vertisols	Aridsols
Annual erosion	Highly restricted	Restricted to mod	Intense	Mod to restricted	Restricted
Bedload	Very low	Low to moderate	Very high	Moderate	Very low
Suspended load	Very low	Low to moderate	Very high	Moderate	Very low
Dissolved load	Very low	Low to moderate	Moderate	High	Low
Continental sedimentary response:					
Siliciclastics	Highly restricted	Restricted	Greatest	Moderate	Highly restricted
Chemical	Domed peats	Planar peat	Planar peat?	Carbonates	Evaporites

From Cecil (1990)

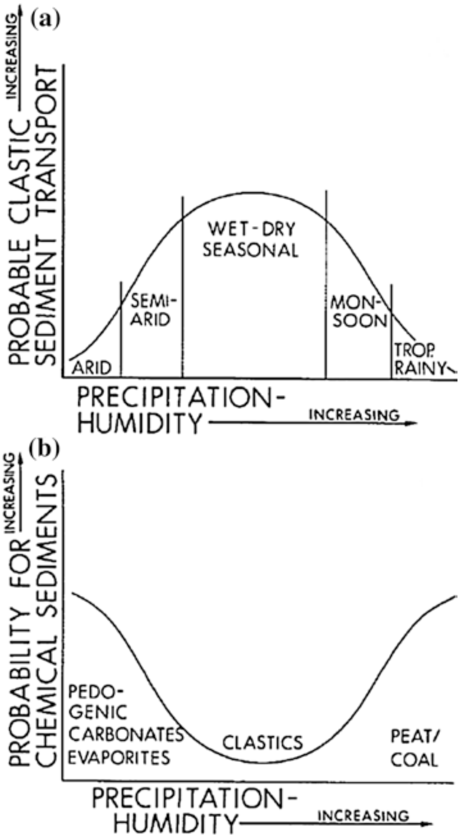


Fig. 5.16 Sedimentological response to climate change. **(a)** Probability for clastic input in response to climatic wetness. **(b)** Conditions for formation of chemical sediments (Cecil 1990, Fig. 1, p. 533)

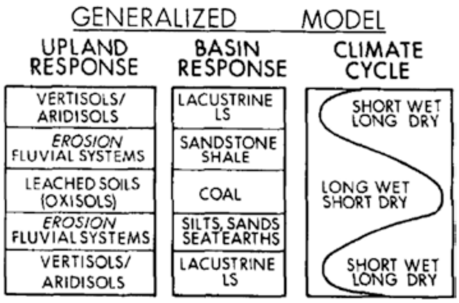


Fig. 5.17 Sedimentary response to changes in paleoclimate in the Pennsylvanian, based on interpretation of cyclothems (Cecil 1990, Fig. 4B, p. 535)

area of grassland, and the rainforest belt a zone within which weathering and erosion processes will tend to generate more suspended chemical sediment load than clastic bedload. Cooler global climates, such as those that characterize glacial episodes, will tend to shift these climate belts southward, meaning that more of the Niger watershed lies within the steppe zone, with consequent increased levels of

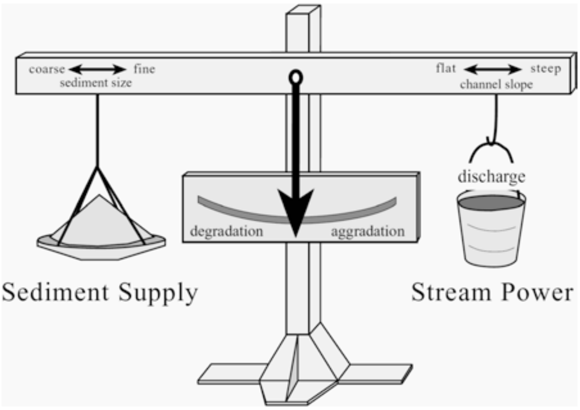


Fig. 5.18 Balance model for aggradation and degradation of alluvial channels, emphasizing changes in the relationship between discharge and sediment supply (Blum and Törnqvist 2000, Fig. 8, p. 12; after Lane 1955)

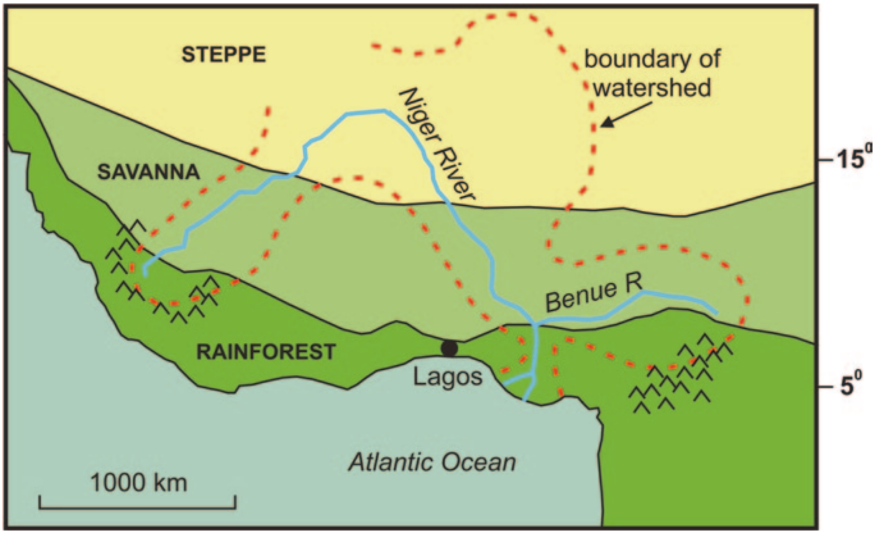


Fig. 5.19 The climate zones of the Niger River watershed. Movement of these belts as a result of climate change can be expected to exert a major control on the amount and type of sediment delivered to the Niger delta (after Van der Zwan 2002)

clastic sediment delivery to the delta. Conversely, during periods of warmer climate, the rainforest belt would be expected to expand northward, with a resulting reduction in the clastic sediment load. These variations might be expected to be present in the form of cyclic variations in sediment calibre or sand-bed thickness in the Niger delta (although this would likely be masked on the delta front because of autogenic overprinting by slumps, turbidite events, etc.).

Figure 5.20 shows a model of fluvial processes in relationship to glacially controlled changes in climate and vegetation, based on Dutch work. These studies, and similar work in Texas (Blum 1993), deal with periglacial regions, where climate change was pronounced, but the areas were not directly affected by glaciation. Vandenberghe (1993) and Vandenberghe et al. (1994) demonstrated that a major period of incision occurred during the transition from cold to warm phases because runoff increased while sediment yield remained low. Vegetation was quickly able to stabilize river banks, reducing sediment delivery, while evapotranspiration remained low, so that the runoff was high. Fluvial styles in aggrading valleys tend to change from braided during glacial phases to meandering during interglacials (Vandenberghe et al. 1994). Vandenberghe (1993) also demonstrated that valley incision tends to occur during the transition from warm to cold phases. Reduced evapotranspiration consequent upon the cooling temperatures occurs while the vegetation cover is still substantial. Therefore runoff increases, while sediment yield remains low. With reduction in vegetation cover as the cold phase becomes established, sediment deliveries increase, and fluvial aggradation is reestablished.

It is apparent that fluvial processes inland and those along the coast may be completely out of phase during the climatic and base-level changes accompanying glacial to interglacial cycles (Fig. 5.20). Within a few tens of kilometres of the sea, valley incision occurs at times of base level lowstand, during cold phases, but the surface may be modified and deepened during the subsequent transgression until it is finally buried. Inland, major erosional bounding surfaces correlate to times of climatic transition, from cold to warm and from warm to cold, that is to say during times of rising and falling sea level, respectively.

Another clear example of climate change affecting fluvial sedimentation was provided by Foreman et al. (2012). The Paleocene-Eocene thermal maximum (PETM) represented a period of about 200 ka during which it has been hypothesized that global climates were substantially warmer than earlier in the Paleocene

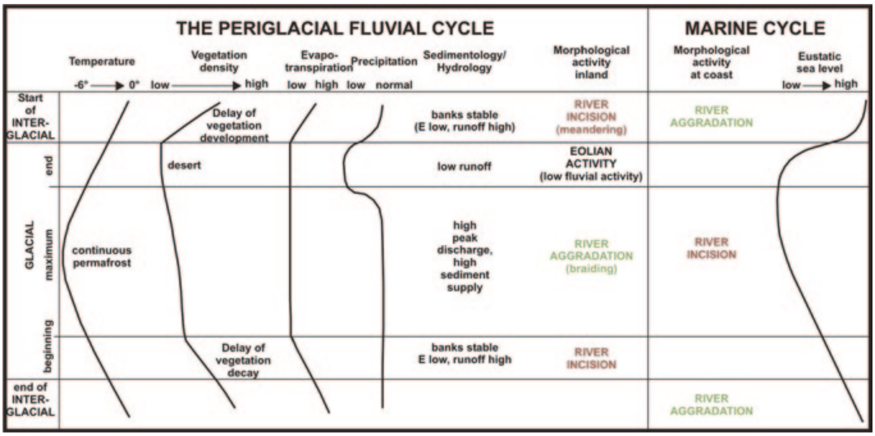


Fig. 5.20 Relationship among temperature, vegetation density, evapotranspiration, precipitation, and sedimentary processes in river systems during glacial and interglacial phases, and the relationship to the contemporaneous marine cycle. Based on work in the modern Rhine-Meuse system (Vandenberghe 1993), with the marine cycle added

and, subsequently, later in the Eocene. The cause is beyond the subject matter of this book, but in one of the nonmarine Laramide Basins of Colorado, the Piceance Creek Basin, it would appear that the climate change imposed a very distinctive change on the fluvial style. An interval of some 40 m of fluvial strata, in the middle of a continuous Paleocene-Eocene section, shows a marked shift from thin, laterally restricted sand bodies, interpreted as the product of small, shallow rivers, to much thicker, broader sand bodies in which the evidence of upper flow-regime plane-bed sedimentation is particularly common and distinctive. Foreman et al. (2012) are able to rule out tectonic or other controls that brought about these changes, and the fact that they can be correlated in time with the PETM is more than suggestive.

Blum (1993) developed a detailed radiocarbon-based chronostratigraphy for several of the rivers draining into the Gulf of Mexico. The alternation of episodes of floodplain aggradation and channel incision records changing climatic conditions since the peak of the last ice age. The chronology of events in the Upper Colorado drainage of central Texas, illustrating the evolution of late Pleistocene through modern alluvial stratigraphic sequences, is summarized in Fig. 5.21.

Diagram (a): During the last full-glacial period, approximately 20–14 ka, sediment supply exceeded transport capacity, resulting in deposition of late Pleistocene fills; (b) approximately 14–11 ka, sediment supply greatly diminished resulting in abandonment of late Pleistocene flood plains and excavation of bedrock valleys; (c) The first of two episodes of aggradation: approximately 11–5 ka, sediment supply exceeds transport capacity resulting in deposition of early to middle Holocene fills; (d) approximately 5–2.5 ka, flood magnitudes decrease resulting in abandonment of early to middle Holocene flood plains and soil formation, but sediment supply remains high, thus promoting storage of sediments and production of unconformity by continued lateral migration of channels; (e) approximately 2.5–1 ka, sediment supply remains high, thus promoting continued lateral migration of channels and storage of sediments, but increases in flood magnitudes result in burial of soil profiles on previously stable surfaces; and (f) last 1000 years, decreases in sediment supply result in clearing of stored sediments, whereas decreases in flood magnitudes result in abandonment of flood plains, and incision of modern narrow valley.

Another example of glacial to interglacial climatic control of sedimentation is provided by the Po Basin in Italy (Amarosi et al. 2008). In this case, the basin is situated near sea level, and the climatic control is expressed by changes in alluvial architecture that reflect the direct control of sea-level. Correlation of laterally amalgamated channel sand bodies in the subsurface is facilitated by the presence of well-developed mud-dominated layers with a characteristic pollen signature of warm-temperate forests (Fig. 5.22). These are interpreted as transgressive deposits corresponding to phases of rising sea-level at the end of a glacial episode. The underlying, laterally amalgamated fluvial-channel bodies "which may be several tens of km in width, are interpreted as complex systems of laterally migrating, braided- and low-sinuosity rivers that developed under conditions of increased sediment supply" during cool, glacial phases. "Low accommodation during lowstand

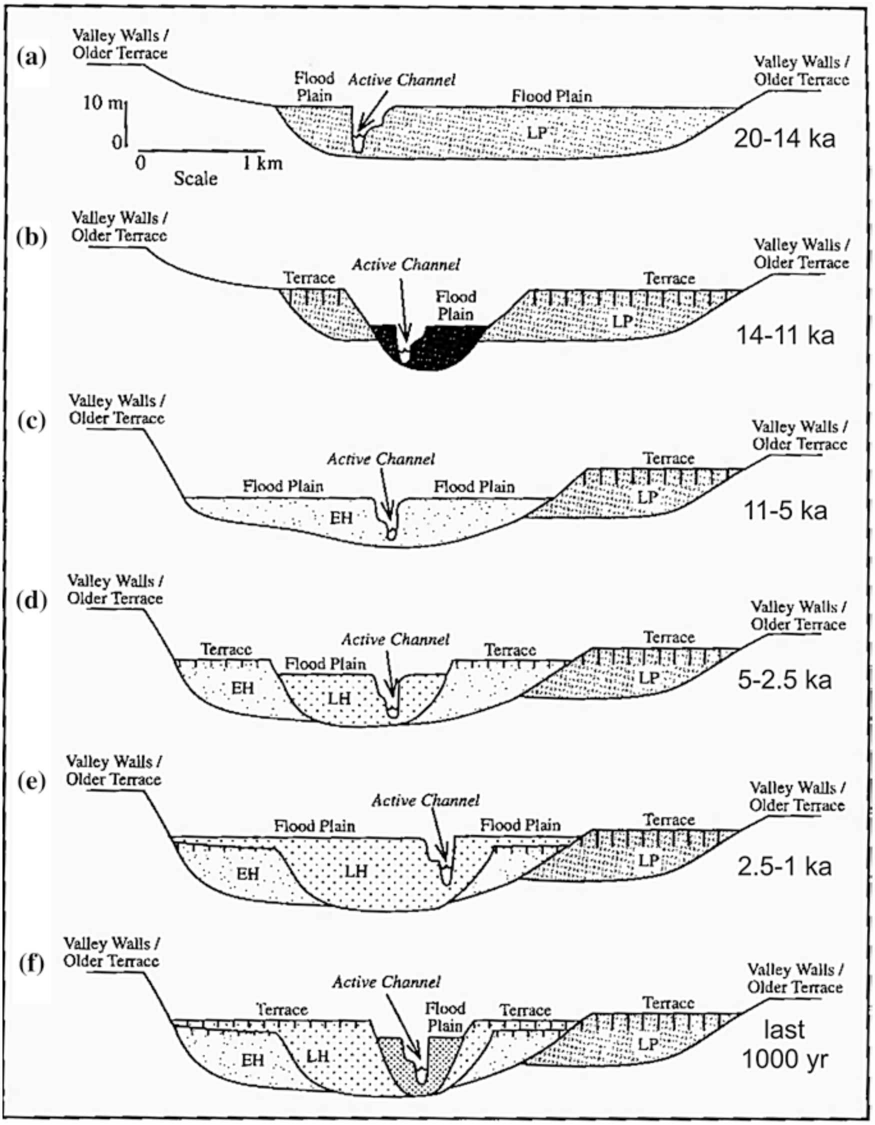


Fig. 5.21 The sequence of events in the Upper Colorado drainage, central Texas. See text for details (Blum 1993, Fig. 6, p. 269). AAPG © 1993. Reprinted by permission of the AAPG whose permission is required for further use

(and early transgressive?) phases favoured lateral migration of river channels, with widespread development of scour-and-fill episodes" (Amarosi et al. 2008, p. 66).

The sharp transitions at cycle boundaries from sheet-like fluvial channel bodies to organic-rich clays and their diagnostic pollen signal (sharp increase in arboreal pollen) may suggest

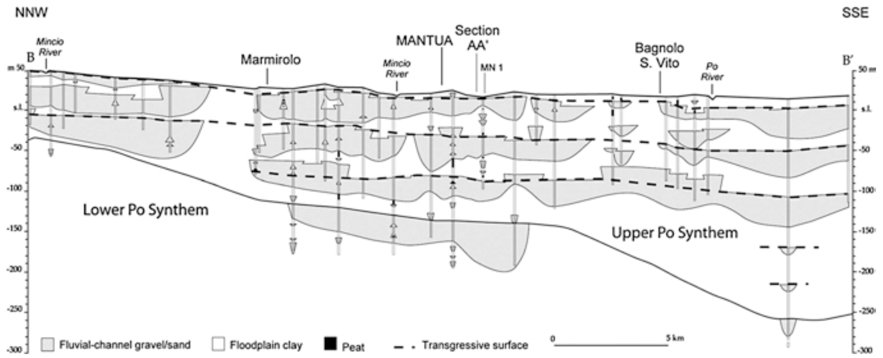


Fig. 5.22 Subsurface Quaternary stratigraphy of part of the Po Basin, Italy. Laterally amalgamated channel sand complexes can be correlated with each other by the presence of distinctive mud-dominated layers (shown by *dashed lines*) characterized by a pollen spectrum typical of warm-temperate forests (Amarosi et al. 2008, Fig. 6, p. 65)

> that generalized development of paludal areas and poorly drained floodplains took place in the study area in response to rapid sea-level rise close to the onset of interglacial periods (TST). During these phases, river channels were probably essentially non-migrating, as suggested by the strongly lenticular, ribbon-shaped channel geometries … Increased accommodation due to the combined effect of subsidence and sea-level rise led to widespread aggradation. (Amarosi et al. 2008, p. 66)

In the Rio Grande Rift of New Mexico, Mack et al. (2011) demonstrated a correlation between phases of terrace development and climate change. Periods of terrace development, including the deposition of floodplain sediments, correlated with episodes of relative humidity, whereas periods of incision occurred during times of aridity. These changes provide another example of the processes illustrated in Fig. 5.18.

In the more ancient clastic sedimentary record climatic control is clearest in the case of lacustrine deposits that have been deposited under the influence of orbital forcing. Two classic examples are discussed briefly here, the Triassic Newark Group of eastern North America, and the Eocene Green River Formation of Wyoming. In both cases, lacustrine cycles, which include evaporites and/or oil shales, are very well developed and have yielded clear orbital signals. Alluvial deposits occur at the margins of the basins, and interfinger with the lake deposits to an extent that varies, depending on the changing climate.

Olsen (1990) defined three types of lacustrine facies complex (Fig. 5.23). Alluvial deposits, consisting mainly of coarse alluvial-fan deposits, are limited largely to the basin margins. Tongues of fluvial sandstone extended further into the lake basin during arid periods, when lake water levels dropped, and ephemeral drainage systems extended across a widening sand flat around the basin margins. A strong orbital signature has been extracted from these rocks (Fig. 5.24). A very similar stratigraphic pattern is illustrated by the Green River Formation

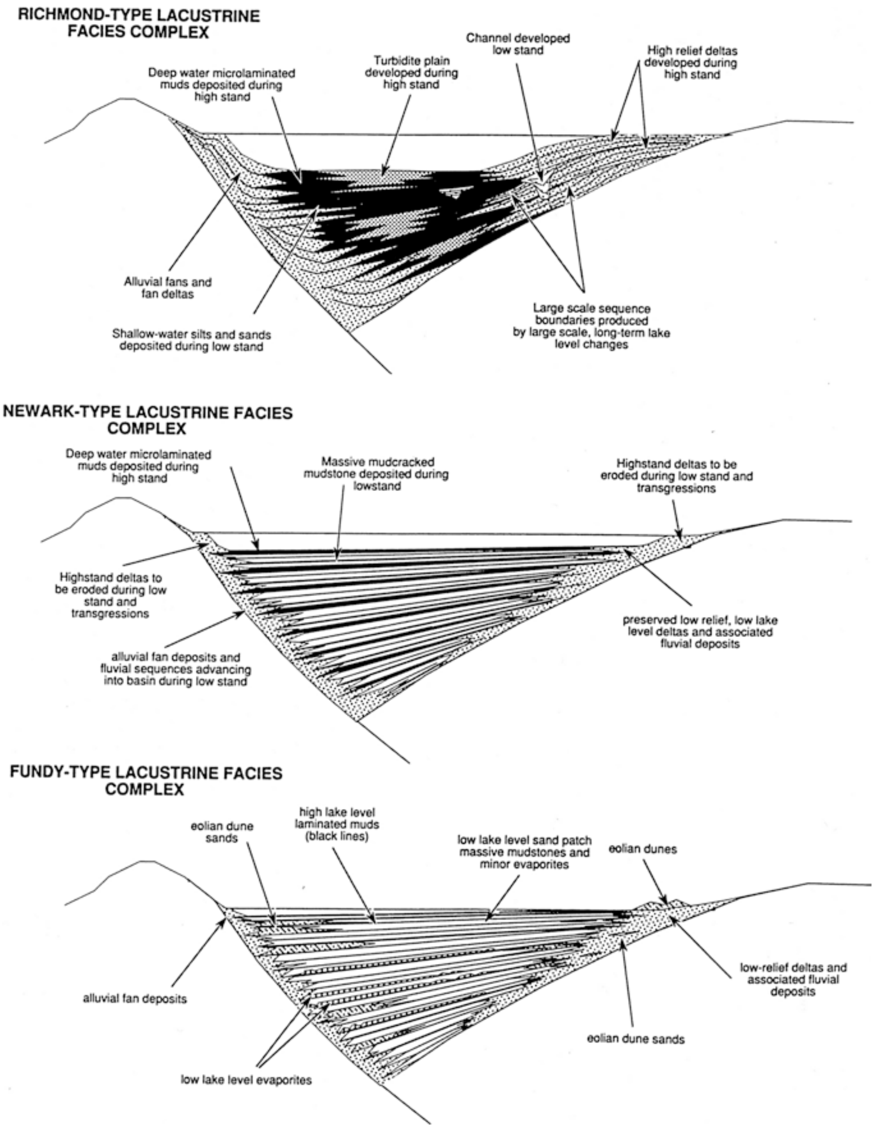


Fig. 5.23 Idealized lacustrine facies complexes, Newark basins of eastern North America (Olsen 1990, Fig. 5, p. 215). AAPG © 1990. Reprinted by permission of the AAPG whose permission is required for further use

(Fig. 5.25). Eugster and Hardie (1975) provided a depositional model which shows how these various facies relate to each other (Fig. 5.26).

Advances in the interpretation of paleosoils, the evolution of vegetation, and the interpretation of facies assemblages and architecture led Allen et al.

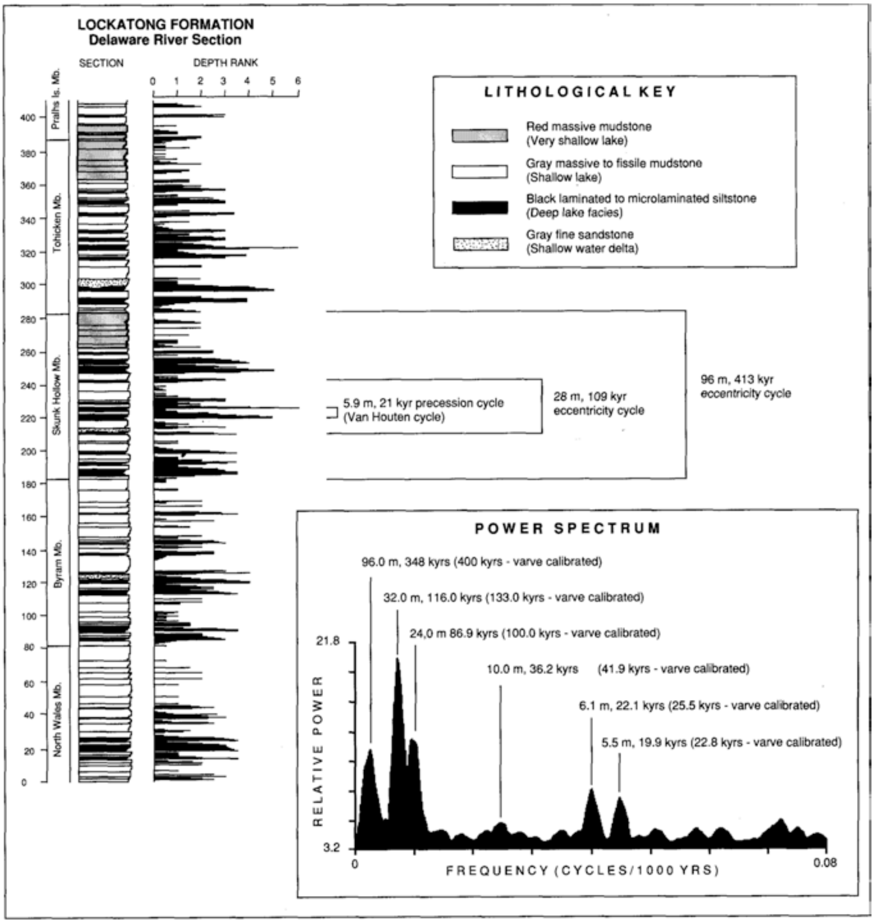


Fig. 5.24 Section of Newark-type lacustrine facies complex in the middle Lockatong Formation, Newark Basin. The power spectrum was derived by Fourier time-series analysis. Depth ranking indicates a ranking of increasing inferred water depth based on sedimentological analysis. Values in kyr are cycle periods in thousands of years (Olsen 1990, Fig. 7, p. 217). AAPG © 1990. Reprinted by permission of the AAPG whose permission is required for further use

(2011) to offer a detailed interpretation of the variations in climate through the Carboniferous to Permian record in Atlantic Canada. This record is largely non-marine, includes thick coals and significant evaporite deposits, and has long been interpreted as the product of deposition on a tropical regime. In a section headed "Fluvial Deposits as Climate Proxies" (p. 1525), they noted that

> Continental records of climate change within the paleotropics have been interpreted using a number of different methods by various researchers, including the spatial and temporal distribution of climate-sensitive lithologies such as coal, evaporites, and eolianite, sequence stratigraphy, paleosols, geochemical proxy analysis, and paleobotanical and paleo-ecological

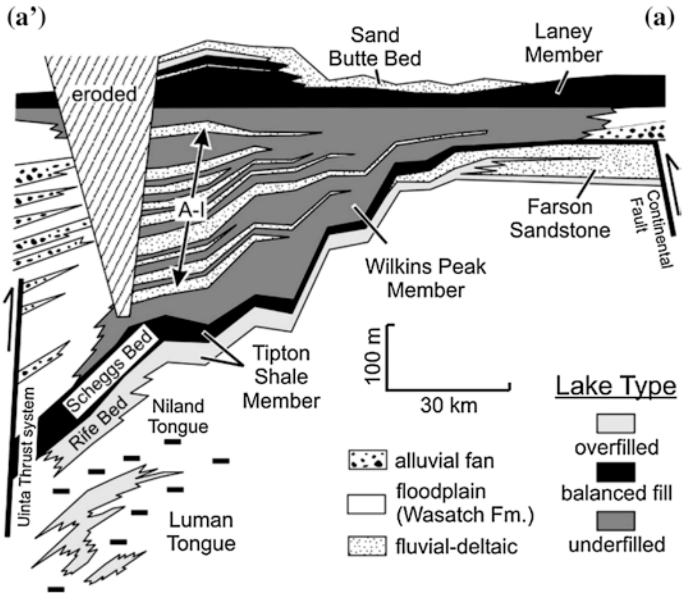


Fig. 5.25 North-south stratigraphic cross-section through the Green River Formation (Pietras and Carroll 2006, Fig. 2, p. 1199)

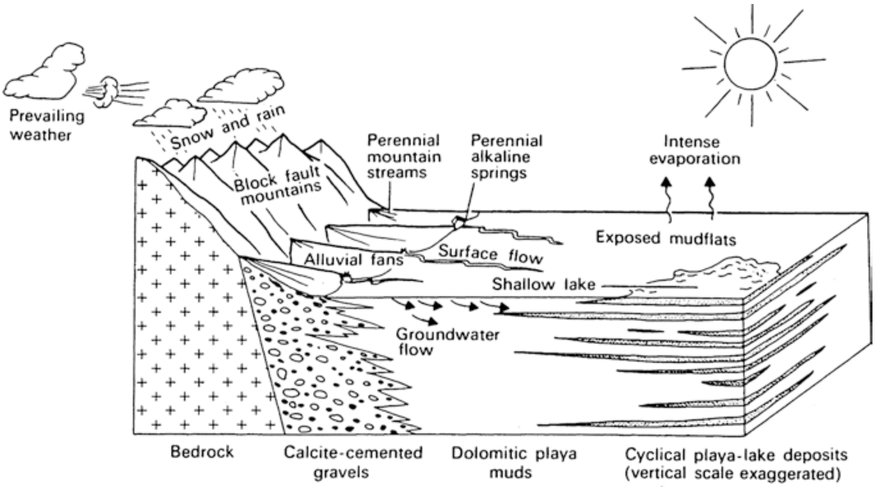


Fig. 5.26 Schematic block diagram showing depositional framework of the Wilkins Peak Member, Green River Formation (Allen and Collinson 1986, Fig. 24, p. 87; after Eugster and Hardie 1975)

analysis. To date, geochemical and paleobotanical records have provided the majority of high-resolution interpretations of paleoclimate change within continental successions.

Amongst the objectives of this paper was a project to employ facies criteria to identify climate change within the fluvial deposits. Earlier studies by the authors had suggested that a distinctive facies association is developed in tropical environments characterized by strongly seasonal discharge (Fielding et al. 2009), and the criteria discussed in this and other papers were applied to the analysis of the fluvial record in the project area. While the results appear internally consistent, it is suggested that it remains the paleobotanical and paleosoil evidence that is the most definitive in identifying climatic conditions in these rocks. These rocks provide a good example of the difficulties in characterizing fluvial style discussed in Sect. 2.2 of this book.

5.2.3 Distinguishing Between Climatic and Tectonic Driving Mechanisms

There are many examples in the ancient record for which speculative interpretations of climate change have been invoked to explain facies variations. In Sect. 3.4 we discuss the deposits described by Martinius (2000), in which a crude cyclicity between facies associations (Fig. 3.16), clearly of allogenic origin, is interpreted as either tectonic or climatic in origin. Hillier et al. (2007) described facies evidence for contrasting climates in the deposition of the Old Red Sandstone of South Wales. It is unclear from the field evidence whether the contrasting climates represent climatic cyclicity during the basin history, or whether the distinctly facies associations were derived from source areas marginal to the basin that were characterized by different local climatic settings. Kallmeier et al. (2010) described coarse fanglomerates deposited by stream flow and sediment gravity flow, with the variation between these facies types likely indicating changing climatic humidity or, possibly, tectonic control.

In this section we use the term *cyclothem* to refer to sequences generated by orbital forcing and the term *cyclothemic* to refer to cycles that may be similar to cyclothems in facies, thickness, and time span, but are not necessarily generated by orbital forcing. There is an overlap in frequency and sedimentation rate between cyclothems, as thus defined, and those attributed to high-frequency tectonism (*SRS-8,-9* and *-10*). Dickinson et al. (1994) noted that there are four types of processes that can potentially generate cyclothemic-type deposits: (1) orbital forcing, (2) autogenic processes, such as delta-lobe switching, (3) movement of faults and folds, and (4) flexural loading of individual thrust plates. Criteria for distinguishing between these various processes are clearly essential. A summary of the comparisons and contrasts between cyclothems and cycles of tectonic origin is presented in Fig. 5.27.

Tectonic cycles	Milankovitch cycles
May constitute an hierarchy of cycles spanning 10^3-10^7-year durations	May constitute an hierarchy of cycles spanning 10^4-10^5-year durations
Cycles correlate with tectonic episodes	No correlation with tectonism
No correlation with climate change	Cycles may be climatic or glacioeustatic. Cycles may contain rhythmic facies variations related to temperature/redox/productivity cycles
May contain internal angular unconformities	Internal angular unconformities not present
Clastic facies cycles correlate with cycle boundaries (f-u and c-u trends consistent with tectonic control)	Clastic and chemical cycles in different parts of a basin may correlate with each other
May show reciprocal vertical facies trends across subsidence/uplift hingelines	No facies relationships to structural trends
Confined to a particular basin or orogen	Cycles could potentially correlate across continent, or globally

Fig. 5.27 Table summarizing the key differences between sequences generated by tectonic mechanisms, and cyclothems generated by orbital forcing

Some autogenic deposits may potentially be confused with true cyclothemic deposits on the basis of vertical profile characteristics. However, autogenic deposits are limited in areal extent to the depositional system which they represent. The largest delta systems (e.g., Mississippi, Niger, Nile) are on the order of about 100 km across. Therefore, the lateral extent of a cyclic unit is clearly an important criterion by which to distinguish autogenic facies successions from regional cycles caused by allogenic mechanisms.

The two major allogenic driving forces that develop cyclothemic deposits on a regional scale are orbital forcing and flexural loading. The immediate effects of flexural processes may be to generate accommodation close to the point of load, giving rise to the classic "lozenge-shaped" regional isopach distribution. However, because the crust has flexural strength and may transmit stress "in-plane" through the crust, flexural effects may be generated over wide areas of a continent. In fact, as discussed elsewhere (Miall 2010, Sect. 10.4), changes in intraplate stress may be capable of generating tectonic events and rapid changes in accommodation on a continental, hemispheric, even, potentially, on a global scale. However, high-frequency sequences that are deposited as a result of the changes in accommodation and sediment supply caused by tectonism will have characteristics that clearly distinguish them from true cyclothems. Cycle thickness and facies will show clear relationships to structural features within a basin. There may be changes in clastic grain-size that can be related to structural features. For example, coarse conglomerates may be cut by and rest on the thrust faults along which movement has generated the accommodation for a tectonic cyclothem (e.g., Fig. 5.8). Cycles generated by tectonism will typically vary in thickness and facies along trends that

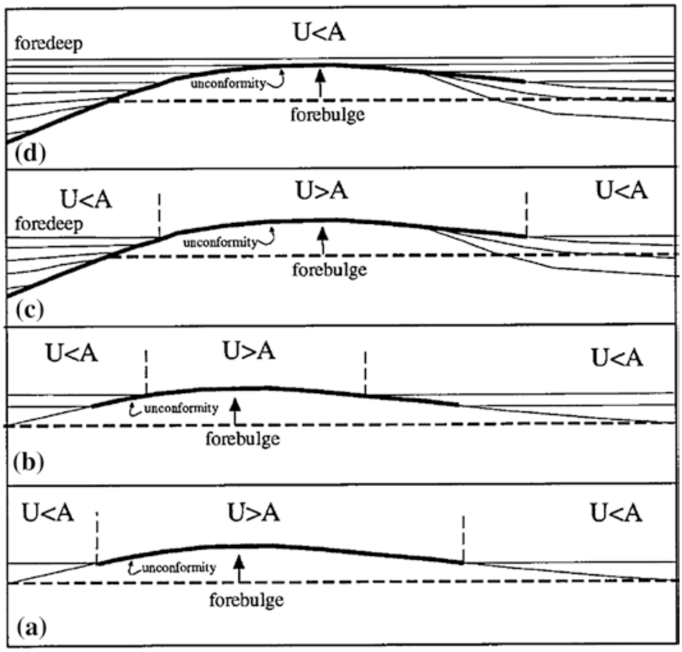


Fig. 5.28 The evolution of a forebulge, showing the development of an unconformity as a result of uplift and cratonward migration, and the onlap of foredeep and back-bulge-basin strata (Currie 1997, Fig. 6, p. 1211)

parallel tectonic grain, and may contain internal architectural features, such as onlap patterns and angular truncations that indicate the syndepositional nature of tectonism (e.g., Figs. 5.9, 5.13, 5.28). Where the proximal regions of basin fills are preserved, that is, the areas where active structures affect contemporaneous sedimentation processes, growth strata may be defined and mapped, which show a clear relationship between sedimentation and tectonics, such as progressive unconformities (e.g., Riba 1976; Anadón et al. 1986; Barrier et al. 2010), The reciprocal -stratigraphy process described by Catuneanu et al. (1997b, 1999, 2000) is a particularly clear example of the way in which tectonism may leave an unmistakable imprint on stratigraphic architecture.

Conversely, cyclothems and other sequences generated by orbital forcing may be expected to extend across tectonic elements, such as forebulges, without major change in thickness, but may change in facies. If eustatic sea-level change, or continental-scale climate change, are the major driving forces in sequence generation, then although sequences may show major internal facies changes, they may, nevertheless, extend across tectonic boundaries and be correlatable across and between sedimentary basins that are the product of a range of tectonic environments. This is the basis for the erection of the Greenhorn cycles of the Western Interior Basin (This unit is present in the lower right part of the section illustrated in Fig. 5.5)

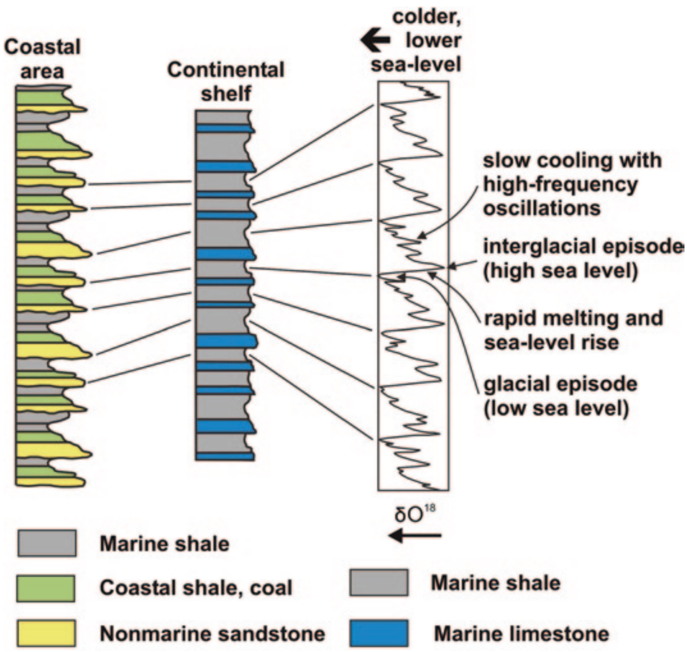


Fig. 5.29 A general model of "non-glacial" Milankovitch cyclicity, based on cycles observed in the Western Interior Seaway. These are correlated against an idealized oxygen isotope curve, showing the relationship between sea level, the alternation of warm and cool global temperatures, and the resulting sedimentary facies. Cooler, wetter episodes, which may correlate to glacioeustatic lows, are periods of higher sediment yield, with the development of significant nonmarine clastic wedges, and delivery of fine clastic material to the continental shelf. During warmer or drier phases clastic supply is limited, coastal deposits are thinner or finer grained, and on the continental shelf biogenic carbonate production may become the dominant sedimentary process. (Generalized from concepts in Elder et al. 1994)

and a general model of non-glacial Milankovitch cycles shown in Fig. 5.29. A cyclic change in climate is the only mechanism that could generate sequences that change facies assemblage entirely from location to location. Likewise, cycles that consist entirely of changes in chemical sedimentation, such as marl-limestone rhythms, can only be the product of climatic forcing of changes in water chemistry or organic productivity.

Dickinson et al. (1994) used these ideas in the construction of some simple but elegant diagrams that clarified the differences between the effects of the two major allogenic driving mechanisms. Figure 5.30 compares the successions in two contemporaneous basins, the Paradox basin in Utah, and the Pedregosa Basin in the Ouachita-Marathon foreland of southeast Arizona. Each column shows 17 cycles. The scale of the cycles is comparable, and it is concluded that they likely correlate, even though the absence of diagnostic fusulinids precludes a definitive

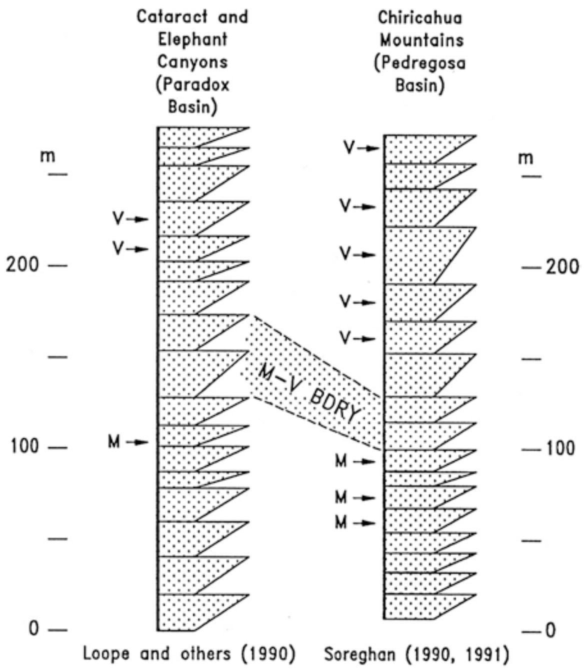


Fig. 5.30 Comparison of cycles deposited in two contemporaneous basins in the (Late Paleozoic) Ancestral Rockies region of the SW United States. *On the left*, cycles of alternating marine carbonate and terrestrial eolianite in the Paradox Basin. *On the right*, shoaling upward carbonate cycles of the Pedregosa Basin. The column style indicates upward decreases in water depth and/or increased subaerial relief, from left to right, in each column (Dickinson et al. 1994, Fig. 6, p. 30)

correlation. As they stated (p. 30): "the very existence of persistent and widespread stratigraphic cycles may well afford the means to achieve stratigraphic correlation throughout the continent at a scale difficult to attempt with confidence using stratigraphic criteria alone."

Dickinson et al. (1994) went on to argue that the style of cycle generated by high-frequency tectonism as a result of flexural loading would be quite different. He compared two successions from foreland basin settings with the Pedregosa basin cyclothems (Fig. 5.31). That in the Antler foreland involve "alternations of oolitic to peloidal grainstones with intertidal to supratidal laminites. They reflect modest variations in water depth, from wave-washed shoals to peritidal lagoonal environments, with the carbonate platform that developed on the flexural forebulge" of the Antler foreland basin (Dickinson et al. 1994, p. 31). The succession is not consistent in facies or thickness across the basin. Contacts between the cycles become gradational as the cycles thicken basinward. The other foreland-basin succession is from the Cretaceous Sevier foreland of the Green River-Tusher Canyon area of central Utah. The deposits consist of interbedded shelf, prodelta, shoreface and delta plain deposits. Again, the cycle frequency and the facies are variable. At least in vertical section, these two foreland basin do not appear to display the regularity that would be expected from deposition under the control of an astronomically precise and regular climatic beat.

Currently there is an interesting debate underway regarding the nature of allogenic controls on parts of the Upper Cretaceous section in the Western Interior

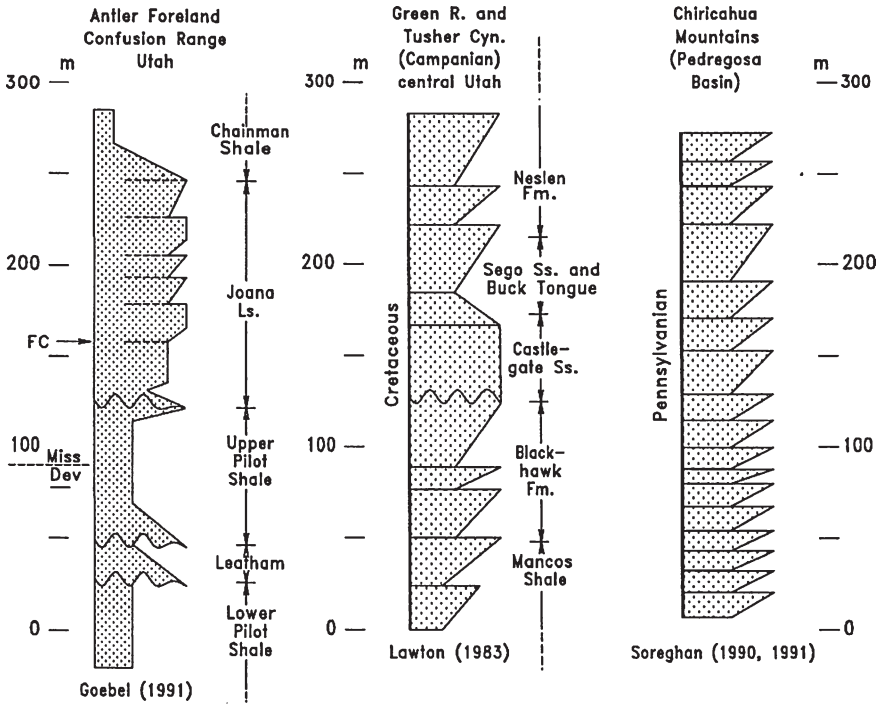


Fig. 5.31 Comparison of the Pedregosa basin succession of cyclothems (*right*; from Fig. 5.30) with two different foreland-basin successions. *On the left*, the Devonian-Mississippian succession of the Antler foreland basin in Nevada; *at centre*, the Upper Cretaceous Sevier foreland basin succession of Utah (Dickinson et al. 1994, Figs. 8, 9, p. 32, 33)

Seaway which, traditionally, has been described as a typical foreland basin setting controlled predominantly by high-frequency tectonism. However, as summarized by Miall (2010, Sect. 11.3.3) isotopic evidence is growing for short-lived glacial episodes during this long period of supposedly "greenhouse" climates. An increasing number of researchers are reporting regional high-frequency cyclicity or rhythmicity in parts of the Cretaceous record of the Seaway, and concluding that the evidence supports the existence of an orbital forcing mechanism, possibly including glacioeustasy (Plint 1991; Elder et al. 1993; Sageman et al. 1997, 1998; Laurin and Sageman 2007; Plint and Kreitner 2007; Varban and Plint 2008). For example, Varban and Plint (2008) pointed to the presence of repeated, widespread regional transgressive bounding surfaces in the Upper Cretaceous stratigraphy of northern Alberta, and estimated "on geometric grounds" that glacioeustatic sea-level changes with amplitudes of around 10 m seemed to be suggested. Antia and Fielding (2011) described the high-frequency cyclicity of the nonmarine-estuarine Dakota Sandstone (Cenomanian-Turonian) in several areas of Utah, and made a case for drawing comparisons to the cyclicity of the contemporaneous

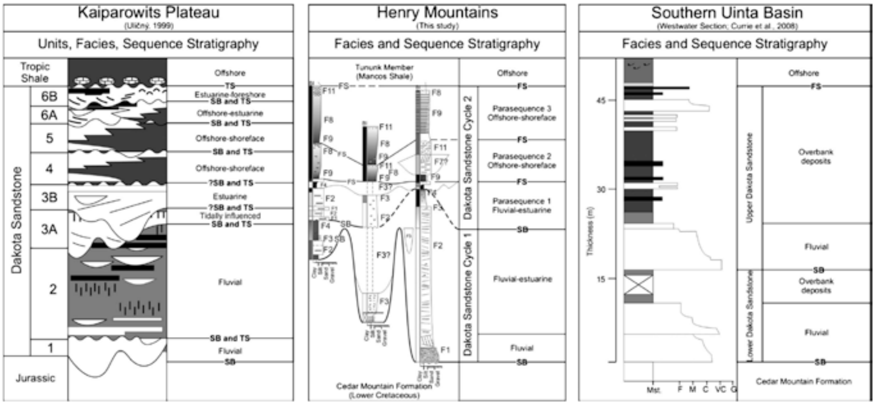


Fig. 5.32 Comparison of facies and sequence-stratigraphic interpretations of the Dakota Sandstone in three areas of Utah. A direct correlation between the cycles in these three sections is not suggested, but the authors point out the similarities between the sections as a possible indicator of influence by high-frequency glacioeustasy (Antia and Fielding 2011, Fig. 16, p. 439). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

coastal-marine to open marine succession in the basin centre further east (Laurin and Sageman 2007). The stratigraphic successions are not rigorously rhythmic (Fig. 5.32) and, clearly, local autogenic processes were active in generating channels, sediment lobes, erosional features, and so on, that complicate the recognition of any possible regional allogenic control. Antia and Fielding (2011) noted the influence of active tectonic features within the basin, and the differences in accommodation between the basin centre, the forebulge and the backbulge; yet, despite these complicating factors, they are led to suggest that low-amplitude but high-frequency eustatic sea-level changes might have played a role in generating the overall stratigraphic architecture. In the absence of clear evidence of tectonism, such as the intraformational angular unconformities described in a different part of the Cretaceous succession by Vakarelov et al. (2006), the orbital forcing model for these deposits appears increasingly possible.

Another complex problem of causation is the cyclicity in the Chinle Formation of the American southwest. This unit, which spans most of the Triassic, has been subdivided into a series of members, all of them consisting of nonmarine clastic deposits representing a wide range of facies and depositional settings. Figure 5.33 illustrates the succession in the area of the Petrified Forest National Park of Arizona, and Fig. 5.34 highlights the major regional unconformities (TR-1, TR-3, J-0) and sequence boundaries (SB 1 to 6) that have been recognized in this unit in Arizona and Utah.

The "major cycles" defined in Fig. 5.34 each correspond to one of the members of the Chinle, and are bounded at top and base by sequence boundaries. The

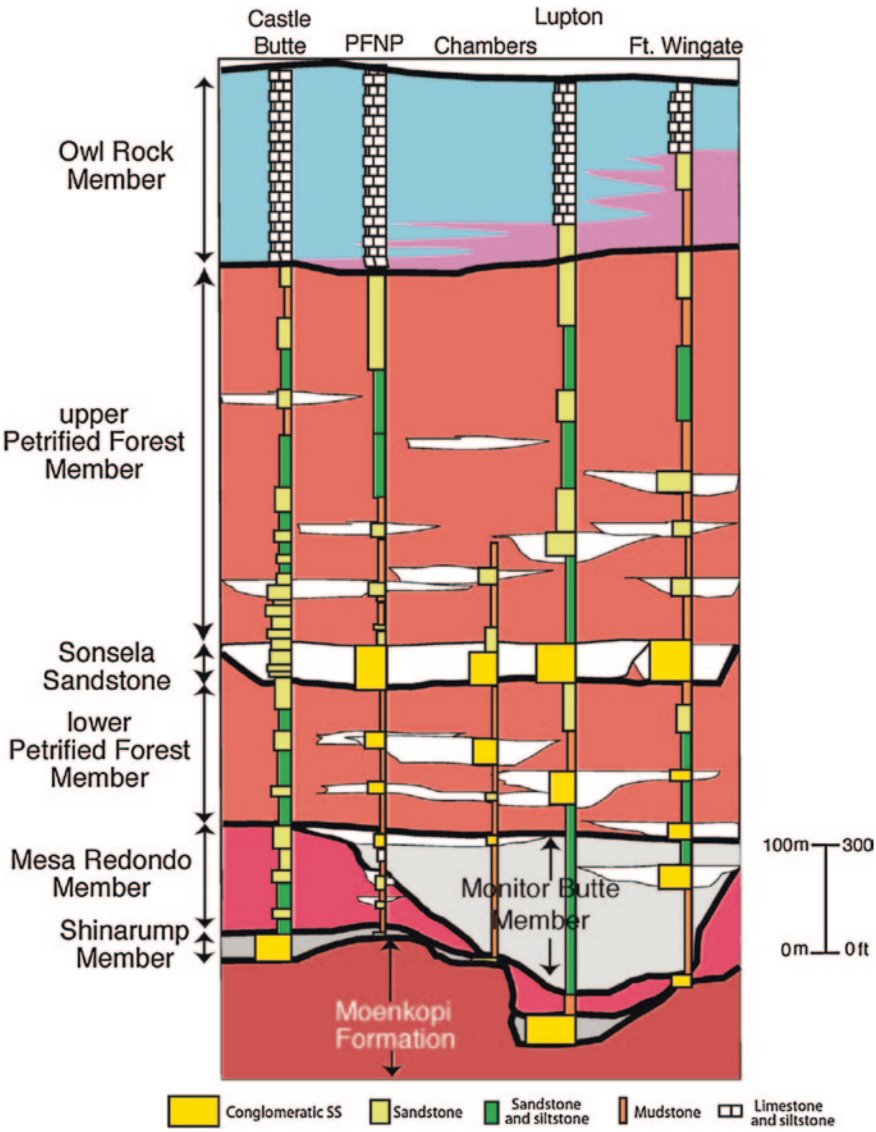


Fig. 5.33 The Chinle formation (Triassic) in the vicinity of Petrified Forest National Park, east-central Arizona, showing the subdivision of the formation into members, the generalized lithology, and the cut-and-fill architecture characterizing the basal members (Dubiel and Hasiotis 2011, Fig. 2B, p. 395)

boundaries are marked by deeply incised paleovalleys and by mature paleosols on interfluve surfaces between the valleys.

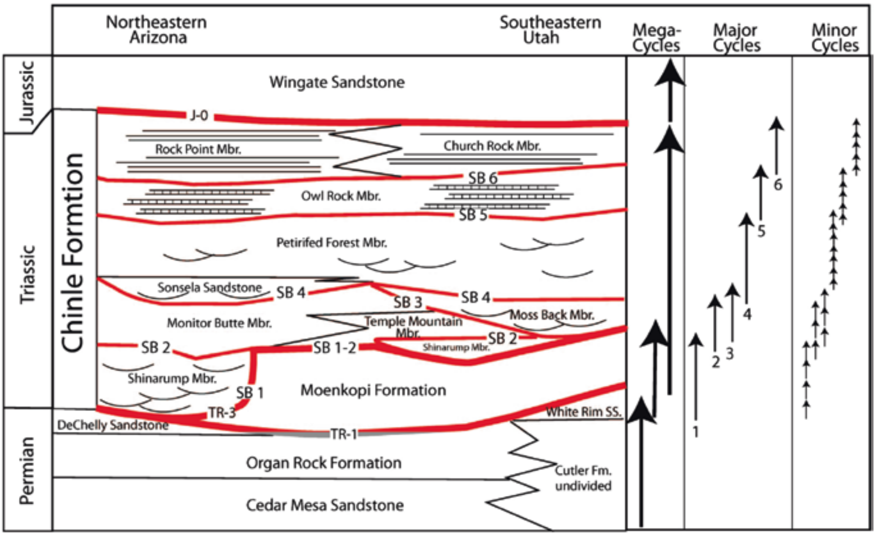


Fig. 5.34 Generalized stratigraphy of the Chinle Formation, Arizona and Utah, showing the major internal unconformities and sequence boundaries, and the interpreted cyclicity. Minor cycles are shown schematically (Dubiel and Hasiotis 2011, Fig. 12, p. 415)

> The progression from valley-fill fluvial systems in the Shinarump Member to a marsh, lake, and delta complex in the Monitor Butte Member and back to paleovalley-fill fluvial systems in the Moss Back Member are examples of the major alternating degradational-aggradational cycles in the lower part of the Chinle. (Dubiel and Hasiotis 2011, p. 412)

These degradational-aggradational cycles were interpreted as climatic in origin, by Dubiel and Hasiotis (2011) but as tectonic in origin by Cleveland et al. (2007). Dubiel and Hasiotis (2011) suggested that the cycles reflect variations in the strength of a monsoon-dominated climate system, citing paleosol and ichnofacies data in support of this interpretation. They also suggested that there was a long-term climatic trend through deposition of the Chinle, from wetter climates and greater landscape stability at the base of the succession, to alternating wet-dry seasonality and greater fluctuations in soil moisture in the middle of the Chinle, to drier climatic conditions with generally low water tables during deposition of the upper Chinle (Fig. 5.35).

Cleveland et al. (2007) described three scales of nested cycles (Fig. 5.36). The definition of these cycles is, in part, based on observations of paleosoil maturity. The smallest scale of cycle, representing facies aggradational cycles (FACs) were interpreted as autogenic channel-fill cycles; the intermediate scale were interpreted as the product of channel avulsion combined with a net drift of the alluvial belt across the valley, resulting in thinner floodplain deposits. The largest scale of cycle, some hundreds of metres in thickness, representing 1–2 m.y., were

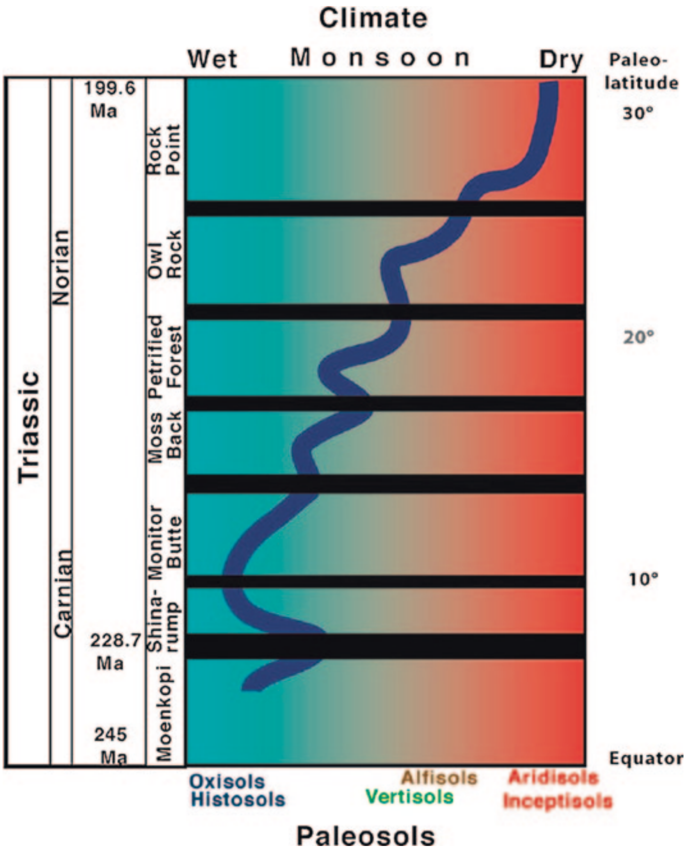


Fig. 5.35 Climatic trends through the Triassic, as evidence in the Chinle Formation of Arizona and Utah (Dubiel and Hasiotis 2011, Fig. 13D, p. 417)

interpreted as the product of pulses of source-area uplift and stream rejuvenation. They noted (Cleveland et al. 2007, p. 922):

> Lucas (1997) and Lucas et al. (1997) describe three-third-order ‘‘sequences’’ in the Chinle strata in New Mexico and Arizona that are attributed to tectonic pulses (Fig. 1). The basal parts of these cycles are characterized by coarse, bedload deposits of multistory channel complexes (e.g., Shinarump Formation of southern Utah and the age equivalent Agua Zarca Formation of northern New Mexico; see Fig. 1). The upper parts of these cycles are characterized by finer-grained overbank deposits and interbedded single-story channel sandstones of suspended-load fluvial systems (e.g., Petrified Forest Fm. and equivalents). Between these cycles are the regional TR-4 and TR-5 unconformity surfaces (Lucas 1997; Lucas et al. 1997). The sequence-scale cyclicity in this study has a higher frequency than that of the ‘‘third-order sequences’’ of Lucas (1997) and Lucas et al. (1997). The ‘‘third-order sequences’’ have a frequency of approximately 4–10 Myr, whereas the fluvial sequences identified herein likely have a frequency closer to 1–2 Myr (Fig. 1). Because eustatic and climatic changes are unlikely, higher-frequency pulses of source area uplift and/or subsidence are the most likely mechanism for sequence-scale deposition of this

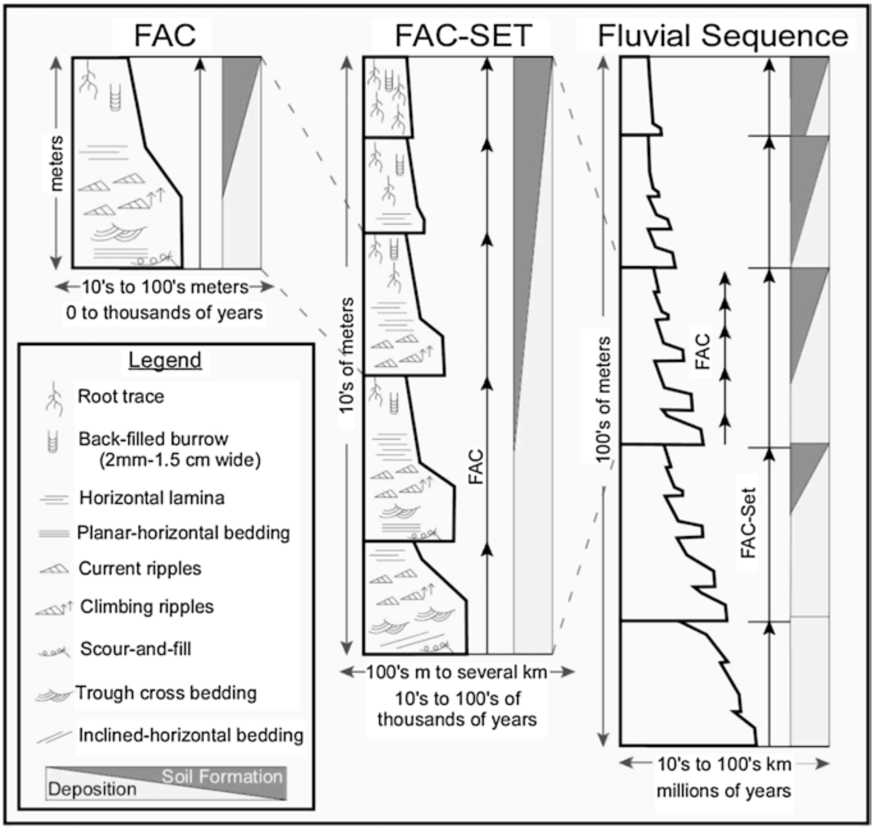


Fig. 5.36 Cycles in the Upper Triassic Chinle Formation of northern New Mexico (Cleveland et al. 2007, Fig. 7, p. 917)

study. Regional-scale tectonism can cause changes in deposition rates and avulsion frequency on a 1 Myr scale, which can result in changes in mean grain size and the proportion of channel deposits in strata on the 100 m scale.

The suggestion of a tectonic mechanism for the larger-scale cyclicity is consistent with many other studies of nonmarine sedimentation where sediment supply and fluvial style are at least in part dependent on tectonically-determined paleoslopes. In this case, there are no obvious indicators of tectonic control, such as changes in paleoslope at the sequence boundaries, whereas the paleosoil and ichnofacies evidence compiled by Dubiel and Hasiotis (2011) seems more consistent with their interpretation of climatic cyclicity. However, comparable cycles on the hundreds-of-metres scale in the Cretaceous-Tertiary section of the Tornillo Basin in West Texas are interpreted as possibly caused by eustatic sea-level changes which, it is suggested, cause changes in the rate of accommodation generation up to several hundred kilometres upstream from the coastline (Atchley et al. 2007). Clearly, each case needs to be examined individually.

5.3 Downstream Controls

Early sequence models, strongly influenced by the work of Vail et al. (1977) and the other major Exxon contributions (Posamentier and Vail 1988; Posamentier et al. 1988; Van Wagoner et al. 1990) focused on eustatic sea-level change as the predominant allogenic control on coastal depositional systems, including eustatic changes driven by global tectonism and changes in sea-floor spreading rates, and glacioeustasy. To these processes must be added relative changes in sea-level caused by local or regional tectonic changes in basin elevation.

It was long assumed that a fall in sea level would lead to incision of river valleys or to extension of the fluvial system to a new, lower river mouth (or both), whereas a rise in sea level would lead to regional transgression, and the flooding of incised valleys, which thereby become estuaries. Based on the early geomorphic work of Gilbert, Davis and others, Mackin (1948) articulated what has long been a fundamental concept regarding the development of fluvial longitudinal profiles: streams grade themselves to base level.

> The graded stream is one in which, over a period of years, slope is delicately adjusted to provide, with available discharge and with prevailing channel characteristics, just the velocity required for the transportation of the load supplied from the drainage basin. The graded stream is a system in equilibrium; its diagnostic characteristic is that any change in

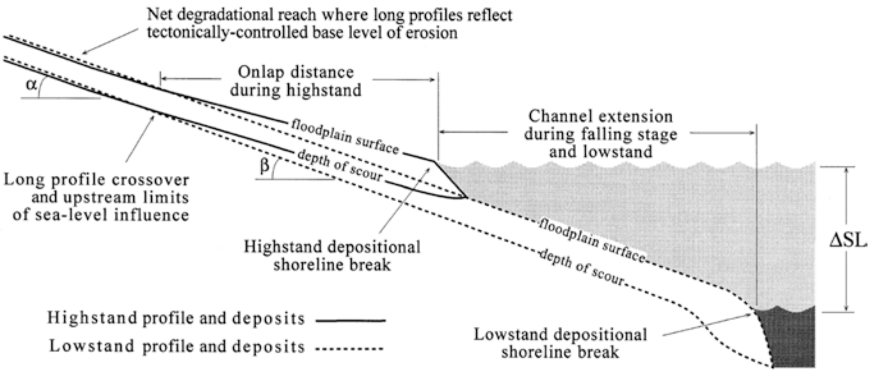


Fig. 5.37 Definition sketch for fluvial response to sea-level change along a continental margin with a distinct highstand depositional shoreline break. Diagram illustrates concepts of channel extension during sea-level fall and lowstand, vs. upstream limits of onlap during sea-level rise and highstand. (figure and caption from Blum and Törnqvist (2000), Fig. 13, p. 18)

> any of the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effect of the change.

Figure 5.37 illustrates the key components of the fluvial response to sea-level change. There has been an extended debate regarding the effect of base-level change on fluvial processes. Early assumptions that a fall in base level would cause rivers simply to incise to create a new point of grade have been shown to be simplistic (Miall 1991b). Similarly, an assumption that incision, beginning at the mouth during a fall in sea level, would simply work its way back upstream as a “knickpoint” (Butcher 1990) have also been shown to be a simplification. There are several degrees of freedom in the physical response of rivers to environmental change. A change in sea level necessarily involves a lateral shift in the position of the mouth of the river. As Miall (1991b) argued, whether a river incises or aggrades during a fall in sea level depends in part on the difference in slope between the original slope of the lower course of the river and the new slope exposed by the sea-level fall. Schumm (1993) demonstrated that to a considerable degree, changes in slope may be accommodated by a change in fluvial style, with little or no change in the balance between aggradation and incision. Leeder and Stewart (1996) added the additional significant factor that much also depends on sediment supply. Where sea-level fall takes place across a gently-sloping shelf, a river bearing a significant sediment load would not incise its valley and may, instead, construct a prograding alluvial plain.

Sequence models for fluvial systems all incorporate the concept of the incised valley as defining the sequence boundary (Wright and Marriott 1993; Shanley and McCabe 1994). However, as discussed in Chap. 6, much is now known about the conditions under which such valleys form, and the significance of these developments for sequence models.

As noted in the previous section, changes in sea level during the late Cenozoic were accompanied by significant changes in sediment supply, which makes it difficult to separate cause and effect in the development of the final preserved stratigraphic record. The stratigraphic generalizations emerging from studies of the valley and terrace record of the Gulf Coast and the Rhine-Meuse system may not be fully applicable to other periods of adjustment in the geological past, when the conditions of change may have been very different. Blum and Törnqvist (2000) provided a review of the evolution of ideas in this area, and conclude (p. 19):

> The fundamental responses to sea-level change appear to be channel extension or shortening, coupled with changes in the elevation of channel bases and floodplain surfaces, in order to keep pace with a shoreline that is advancing or retreating and changing its elevation. All other adjustments take place within this context, and should be regarded as nondeterministic, since they depend on alluvial valley, coastal plain, shoreface and shelf gradients …, discharge and sediment supply from the drainage network and local physiographical factors.

As an example of how specific geologic scenarios may evolve in ways that appear counterintuitive, Blum and Törnqvist (2000) cite the case of the coastline of South Island, New Zealand. The Canterbury Plains constitute the deposits of an active braidplain, consisting of the coarse, predominantly gravel deposits eroded from the actively rising Alpine Mountains to the west. This scenario might be expected to have led to the construction of an actively prograding coastline. However, sea-level has been rising steadily along the coastline, which might be expected to create a condition in which accommodation and supply were in approximate balance, generating an actively aggrading coastline. Neither condition prevails. Because of high wave energy much of the coastline is undergoing erosion, with the development of a wave-cut ravinement surface and an actively retreating cliff line (Leckie 1994).

Another issue of considerable significance to the development of appropriate sequence models for coastal fluvial systems is the extent to which sea-level change affects the patterns of incision and aggradation inland from the river mouth. Early ideas on this topic were strongly influenced by the work of Fisk (1944) on the lower Mississippi Valley, who argued that that eustatic control of the river extended some 1,000 km upstream. This view was challenged in later studies, by Saucier (1994, 1996), who concluded that the influence of sea-level change was somewhat less, arguing that the Mississippi was only affected by eustasy in the stretch downstream from Natchez, a distance of about 400 km. Blum and Törnqvist (2000, p. 19 and Table 1) defined the limit of eustatic influence as the "the upstream extent of coastal onlap due to sea-level rise" (see Fig. 5.37) and, drawing on a range of studies, concluded that this limit is highly variable, ranging from 300–400 km for large, low-gradient systems such as the Mississippi, to as little as 40 km for smaller rivers with steeper gradients.

Current research in this area focuses on the recognition of a distinct area, called the *backwater zone*, which is transitional between the nonmarine and marine environments. The upstream limit of this zone is where the channel bed drops below sea level. Bedload transport and deposition drop markedly as the river enters the backwater, and in deltaic settings, avulsions tend to occur more frequently in this zone (Chatananavet et al. 2012). Studies of the ancient record suggest that the transition into the backwater zone can be recognized by facies changes. In the Castlegate Sandstone (Upper Cretaceous) of Utah, it has been postulated that the transition is represented by a downstream change from the fully nonmarine record of amalgamated channels sandstones to a zone of isolated, mud-filled channels with thin overbank deposits (Petter 2011).

Chapter 6
Sequence Stratigraphy

Sequence stratigraphy was developed initially for shallow-marine deposits (Vail et al. 1977), but was extended to the interpretation of coastal fluvial deposits in the models of Posamentier and Vail (1988) and Posamentier et al. (1988). These models included a number of controversial features regarding the response of fluvial systems to base-level change that were discussed at length by Miall (1991, 1996); a discussion not repeated here. Two models developed specifically for fluvial systems were proposed in the early 1990s, those of Wright and Marriott (1993) and Shanley and McCabe (1994), referred to hereafter as the WMSM models. These two papers drew extensively on the concepts relating fluvial architecture to accommodation that had been explored in the numerical simulation model of Bridge and Leeder (1979). That model, in turn, drew on concepts developed earlier by J. R. L. Allen, and was further developed in a series of papers referred to collectively as the LAB models, after the authors Leeder, Allen and Bridge (see Sect. 3.6). However, the LAB and WMSM models are based primarily on sedimentary processes studied in modern rivers and the post-glacial record. As argued in Sect. 6.2, because such studies are focused on relatively high-frequency processes and short time scales, typically at *SRS 7* or *8*, they are of limited value in interpretations of the ancient rock record, most studies of which have been carried out at *SRS 9*, *10* or *11*. This requires a re-evaluation of much of the current research on the ancient fluvial record, a discussion of which is introduced in Sect. 6.3.

6.1 Standard Sequence Models

6.1.1 Development of the Relationship Between Architecture and Accommodation

Studies of alluvial architecture began with the classic work of J. R. L. Allen on the Devonian Old Red Sandstone of Wales (Allen 1974). Allen's original project

A. Miall, *Fluvial Depositional Systems*, Springer Geology,
DOI: 10.1007/978-3-319-00666-6_6,

was to explain the architecture of cycles that included thick and laterally extensive pedogenic carbonate units, which indicated extensive periods of soil development on interfluves. His models postulated cyclic base-level change, incised valleys, or periods when the main feeder stream combed far away across the alluvial plain. Based on qualitative, deductive reasoning Allen (1974) developed five distinct models, with additional variations bringing the total to eight. These represent essentially abstract models in which an interpretation of the Old Red Sandstone was an almost incidental byproduct of the analysis. One of the models is illustrated in Fig. 3.27 as part of a discussion about the controls on avulsion. Another is reproduced here because of its historical interest (Fig. 6.1). The first represents a purely autogenic model, and one which has now been produced by computer simulation, as discussed below and in Sect. 3.6. The second, that shown here, represents, in effect, the first model of fluvial sequence stratigraphy.

In his subsequent papers, Allen turned to the question of sand body connectivity, an issue of obvious relevance to the reservoir characteristics of these successions. The independent variables on which attention focused, and which became the basis for the first quantitative models, were primarily subsidence rate, channel dimensions, and channel avulsion rate. Pedogenic carbonates and exposed interfluves did not feature in any of the subsequent quantitative models discussed below. Indeed, and perhaps ironically, the question became inverted. Instead of exploring the causes of pauses in sediment accumulation, the question became focused on accommodation: how do changes in accommodation affect alluvial architecture?

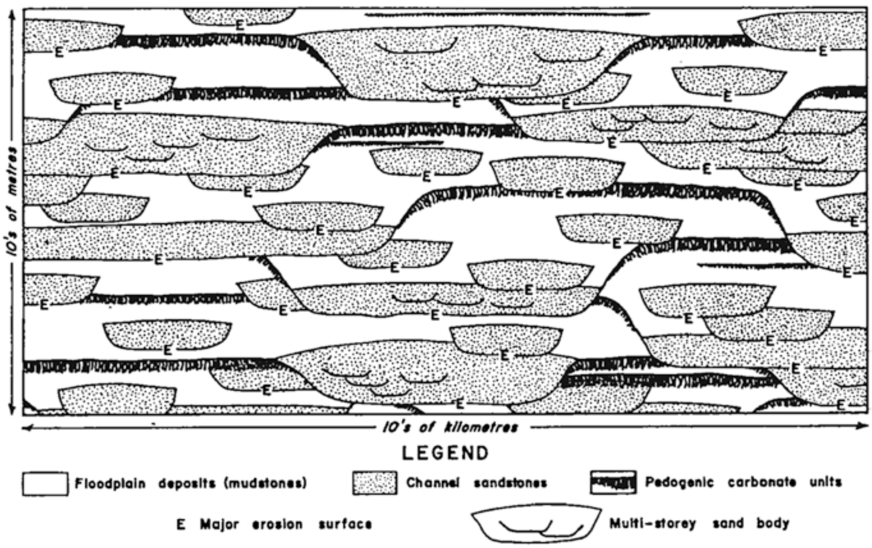


Fig. 6.1 An example of Allen's qualitative models of alluvial stratigraphy. Model 5B, developed by a combination of autogenic channel avulsion and a cycle of dissection and aggradation, in which base level changes over a vertical interval greater than channel depth (Allen 1974, Fig. 9, p. 200)

Allen's speculative architecture models became the basis for a series of quantitative studies collectively termed the "LAB models" after the authors, M. R. Leeder, J. R. L. Allen, and J. S. Bridge (Allen 1978, 1979; Leeder 1978; Bridge and Leeder 1979; Bridge and Mackey 1993a, b; Mackey and Bridge 1995). These studies, in turn, have been examined and experimented with by a number of other researchers (e.g., Heller and Paola 1996; Marriott 1999).

The generation of alluvial stratigraphy models by computer simulation was pioneered by Bridge and Leeder (1979). This is a much-quoted paper (see also discussion of avulsion in Sect. 3.6) that was built on an analysis of the mechanics of channel-belt and floodplain construction and a systematic documentation of the rates and scales of various autogenic sedimentary processes. The main purpose of the model is to build cross-sections of alluvial stratigraphy in an orientation perpendicular to paleoslope. The model is quantitative, in that magnitudes, rates, and scales are stated, and many of the parameters are designed as variable input for successive computer runs to permit an exploration of the complex interaction between dependent and independent variables. Channel-belt sandbodies are assumed to be constructed by any appropriate fluvial style, following the assumption that the two major styles, meandering and braided (the "mobile channels" of Friend (1983)), both construct complexes of bars and minor channels within relatively stable meander belts. This is certainly the case for meandering systems, in which lateral migration is constrained by the alluvial topography of levees and the resistance met by channels when they erode laterally into the fine-grained deposits of abandoned channels and the floodplain. The assumption of a confined channel belt may not be so appropriate for braided systems.

The alluvial stratigraphy models discussed here are autogenic, in that the external, allogenic variables that control the long-term behaviour of the river system are set at constant background values. Tectonism and sea-level change are not assumed to affect the system, so that average accumulation rates and regional paleoslope remain constant throughout a model run. Climate change, and its effect on discharge, fluvial style and sediment supply is also not considered. Bridge and Leeder (1979) were amongst the first to consider the quantitative effects of tectonism on fluvial aggradation, by building in the ability in the computer model to tilt the depositional surface in a direction perpendicular to paleoslope (not discussed here. See Miall 1996, Sect. 11.2.1). The construction of their model is discussed in Sect. 3.6.

Two examples of Bridge and Leeder's (1979) model are illustrated in this book. The most realistic of these is the first (Fig. 3.29), in which sandbody distribution appears to be relatively random after the initial ten avulsion events. The second model (Fig. 6.2) shows a much too regular stacking of the channel bodies, an effect that occurs when channel width and depth are set too large relative to aggradation rate. The effects of the stacking and compaction of numerous preceding events are downplayed in this model, and successive channel positions are determined primarily by the position of the two or three immediately preceding channels. The result is a distribution of channel bodies not unlike that of bricks in a wall. Nevertheless the Bridge and Leeder (1979) model has been much used, particularly by petroleum geologists, because of the insights it yields into channel stacking patterns

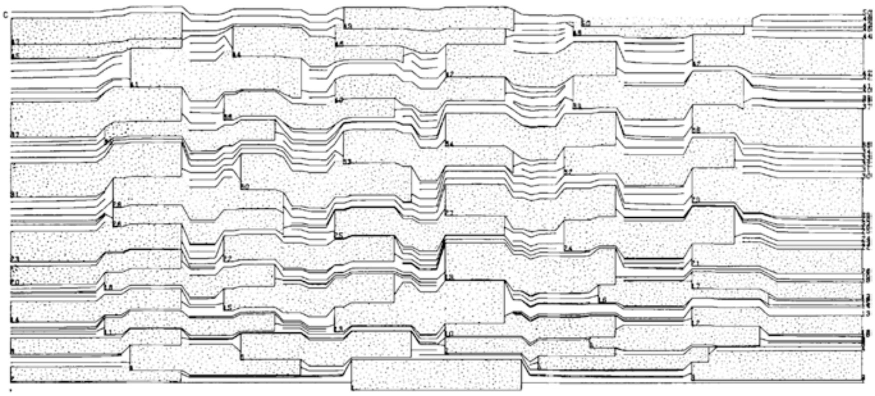


Fig. 6.2 Example of a simulated cross-section of alluvial stratigraphy constructed using the numerical model of Bridge and Leeder (1979, Fig. 2, p. 631). Channel units are set at 2 km wide and 7 m deep

and interconnectedness, factors of considerable importance in the understanding of reservoir predictability and fluid migration behaviour. The model was expanded and updated by Bridge and Mackey (1993a, b), Mackey and Bridge (1995), but the improvements are at a level of refinement that does not need to be addressed here.

According to Heller and Paola (1996, p. 297) "The link between sedimentation rate and channel stacking architecture in the LAB model was a major conceptual breakthrough." These authors went on to note that up to the time of publication, there had been few direct tests of the LAB model because of a lack of information on sedimentation rates and other processes, and their paper also provided a discussion of other controls on alluvial architecture beyond the autogenic processes that constitute the LAB model, such as the importance of different styles of avulsion (nodal versus regional) and the influence of cross-valley and down-valley tilts. Very few of the early studies listed in this section referred to the pioneering work of Wescott (1993), who demonstrated the "complex response" of rivers to forcing processes. His paper focused primarily on the response of fluvial systems to changing base level, but the discussion of the complex way in which rivers adjust to autogenic and allogenic processes has not been widely taken up in subsequent work, apart from a lengthy debate about the formation of incised valleys and subaerial erosion surfaces (Miall 1996, Sect. 11.2.2).

6.1.2 The First Sequence Models for Fluvial Deposits

The issue of accommodation is central to sequence stratigraphy, and concepts emerging from the LAB models were readily incorporated into the first two major sequence models for nonmarine systems, those of Wright and Marriott (1993) and

Shanley and McCabe (1994), referred to as the WMSM models throughout this chapter.

Wright and Marriott (1993) developed a sequence model which they specifically compared with Allen's (1974) Model 5B (Fig. 6.3; compare to Fig. 6.1). The sequence boundary consists of a subaerial erosion surface, into which is cut an incised valley, formed during the preceding falling stage. The erosion surface may be marked by paleosoils, the maturity of which reflects the duration of exposure. The incised valley may include marginal terraces, with preserved fluvial channel or overbank remnants deposited during the falling stage. They suggested that there may be a lowstand deposit characterized by coarse deposits, dominated by amalgamated-channel facies. A transgressive phase is then formed as sea-level begins to rise, with the proportion of coarse channel to fine floodplain deposits depending in part on the rate of sea-level rise. A rapid rise, meaning a high rate of accommodation generation, is presumed to lead to a higher preservation potential for thick floodplain deposits, with isolated channel bodies, and a greater probability for the formation of soils of hydromorphic type, that is, soils formed under saturated, commonly anaerobic conditions. Siderites are common. As sea-level rise slows, in the highstand phase, it is suggested that the rate of accommodation generation decreases, and the rivers evolve primarily by lateral migration and point-bar accretion, leading to the formation of laterally-amalgamated sand bodies, with a much lower proportion of floodplain fines, and a greater probability of the development of mature soils. They then suggested that a subsequent fall in sea-level, which could have the effect of steepening the slope, might be marked by the development of braided systems and the deposition of coarser deposits, although this phase of the cycle would have a low preservation potential because of the high probability of erosional removal as the rivers cut down to a new lowstand.

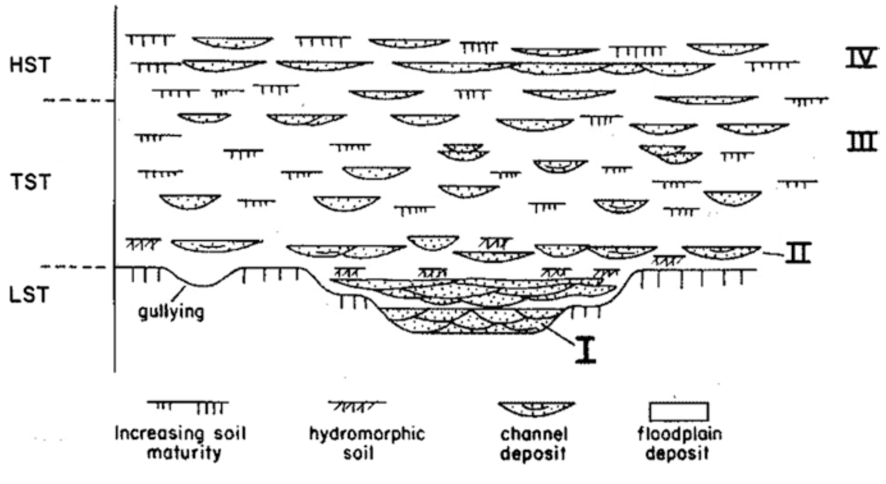


Fig. 6.3 The fluvial sequence model of Wright and Marriott (1993, Fig. 1, p. 206)

Wright and Marriott (1993) provided a few examples of fluvial systems, to illustrate this model, but emphasized that it was largely speculative and hypothetical. However, it has been widely referenced by subsequent fluvial researchers.

The Shanley and McCabe (1994) model was based in part on the authors' earlier study of Upper Cretaceous sedimentation on the Kaiparowits Plateau of south-central Utah (Shanley and McCabe 1989). They emphasized (Shanley and McCabe 1994, p. 546) that "The evolution of stratigraphic architecture at the scale of depositional sequences is governed by the rates at which accommodation space is created or destroyed as well as the sedimentary processes inherent to depositional systems. Interpretation of stratigraphy and application of analogs must reflect an understanding of these fundamental controls." Their summary model (Fig. 6.4) is similar to that of Wright and Marriott (1993), and includes the definition of systems tracts based on the dependence of the alluvial architecture on the rate of accommodation generation. The major difference from the Wright and Marriott model is a discussion of the evidence that might be preserved in a fluvial system of a condition corresponding to the marine maximum flooding surface.

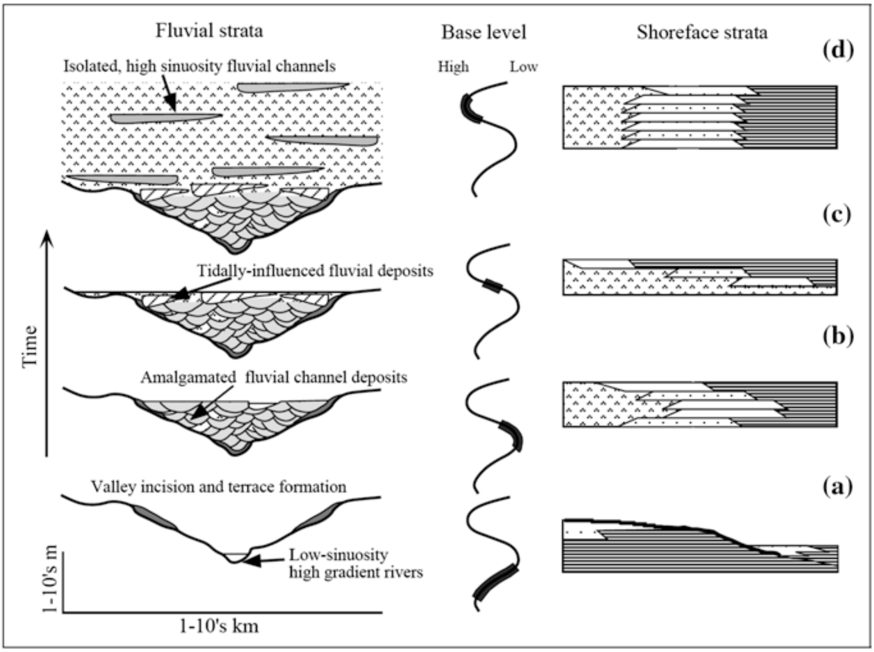


Fig. 6.4 The fluvial sequence model of Shanley and McCabe (1994, Fig. 10, p. 560), showing the relationship between shoreface and fluvial architecture and baselevel change. **a** Falling stage systems tract, with development of incised valley and fluvial terraces, **b** Lowstand systems tract, **c** Tidal influence indicates the beginning of the transgressive systems tract, **d** Highstand systems tract. AAPG © 1994. Reprinted by permission of the AAPG whose permission is required for further use

They suggested that retrogradation of coastal systems during the transgressive phase might lead to "the invasion of tidal processes into areas formerly dominated by purely fluvial processes" (p. 559). In their studies in Utah they documented tidal influence on fluvial strata as much as 65 km inland from coeval shoreline deposits (Shanley et al. 1992).

In both models there is a discussion of how the magnitude of a base-level rise controls whether coastal fluvial deposition spreads out beyond the margins of an incised valley system. Shanley and McCabe (1994) also devoted considerable space to a review of the work of M. Blum on Pleistocene-Holocene fluvial sedimentation along the coast of the Gulf of Mexico, where it has been clearly demonstrated that episodes of aggradation and degradation relate not to sea-level change but to climate change (this work, and comparable studies in the Rhine-Meuse system are discussed at length in Miall 1996, Chap. 13). They offered a caution that

> Our Quaternary models may be superb analogs for the Carboniferous, also thought to be a period of widespread glaciation, but how appropriate are they for the middle Cretaceous, a period of limited glaciation? Application of Quaternary models should be done with at least a modicum of restraint. (Shanley and McCabe 1994, p. 557)

This last caution has not been followed, as demonstrated at length in the next section.

6.2 Rates of Accommodation and Sedimentation

6.2.1 Accommodation and Channel Stacking Pattern

In the LAB and WMSM models, and in subsequent research on nonmarine sequence stratigraphy, it became a central assumption that changes in accommodation rate are critical in the control of alluvial architecture. For example, Wright and Marriott (1993, p. 206) stated of the transgressive systems tract (TST): "During the early stages of the TST the rate of accommodation space creation will be low, producing multistorey sand-bodies," and later in the transgressive phase "Increased accommodation rates favour high levels of storage of floodplain sediments resulting in isolated channels." Other workers, noting the inappropriateness of using terms for nonmarine systems that refer to sea-level changes, suggested terminology for systems tracts such as "high accommodation" and "low accommodation" (Olsen et al. 1995) and "aggradational" and "degradational" (Currie 1997).

Channel-stacking patterns are the basis for the differentiation between high- and low-accommodation systems tracts. During high accommodation, the argument is that channel bodies become buried before the active channel returns to erode or re-occupy the channel site on the alluvial plain. During low accommodation, channels comb through their own deposits and are more likely to develop amalgamated, laterally offset, sand bodies. Channel return depends on the pattern of avulsion.

The condition of high accommodation is defined as a time during which, in some locations, floodplain sedimentation proceeds through at least several channel cycles without evidence of a channel return. The presence of a paleosoil of any level of maturity indicates low accommodation, whereas levees, crevasse plays or pond deposits all indicate normal floodplain accumulation. The distinction between high and low accommodation is therefore a question of the balance between the rate of subsidence/sedimentation and the rate of avulsion. High accommodation may be defined as occurring when channels become completely buried before a channel or meander belt returns to the same location. This will occur whenever $(R_s * a * r) > h$, where R_s = sedimentation rate in m/yr, a = avulsion frequency, r = the number of avulsion events which occur before the channel/meander belt returns to the same location on the floodplain, and h = channel depth. The LAB models use 1000 yrs as a typical avulsion frequency, whereas Stouthamer and Berendsen (2007) demonstrate that in the Rhine-Meuse system the frequency is about 500 yrs. If we set $h = 5$ m, $a = 1000$, and $r = 5$, for continuous floodplain sedimentation to occur, R_S must be > 0.001 m/yr (1 m/ka = 10^0 m/ka). An extreme case may be imagined, where it takes 10 avulsion events (10,000 years) for a channel return to occur, in which case the sedimentation rate must be > 0.5 m/ka (10^{-1} m/ka). These rates and time scales correspond to *SRS 7*. These simple calculations are intended only to help make the point that there is a substantial difference, in most cases, between the sedimentation rates measured in modern and Recent sediments, and those recorded from the ancient record, a point discussed in detail in Sect. 6.2.3. As Heller and Paola (1996) and others have demonstrated, there are several other factors, including the style of avulsion (regional versus nodal) and tectonic tilting of the valley floor that affect channel stacking patterns.

The LAB models were based on a limited range of data derived from a few studies of modern processes, and the post-glacial (post 8000 BP) sedimentary record. Bridge and Leeder (1979) based their models on an average channel-belt aggradation rate of 20 m/ka (10^1 m/ka). All the subsequent LAB studies refer back to this source. Since little of the actual physics of sedimentation can be simulated, these are exercises in dynamic geometry rather than stratigraphy. The WMSM sequence models make some reference to the LAB models, but are largely conceptual. Shanley and McCabe (1994) speculated about the possibly inappropriate use of late Cenozoic analogs as a basis for interpreting the ancient record, but their caution has been ignored. Likewise, Holbrook (2001, p. 216) stated: "The implicit assumption in fluvial LAB-model applications is that channel belts consistently stack in an upward trend to fill accommodation space as it is created. …. It is a reasonable assumption, however, that fluvial sections recording several punctuated episodes of localized valley and nested valley incision, instead of unidirectional aggradation, are common."

The few cases where quantitative stratigraphic data are available from the ancient record reveal some problems with the assumptions behind the models, and it is the purpose of this section to address these.

6.2.2 Time Scales and Sedimentation Rates

Sadler (1981) demonstrated that there is an inverse, log-normal relationship in the stratigraphic record between sedimentation rate and the time span over which it is measured. Sedimentation rates measured in modern settings and the geological record vary by eleven orders of magnitude, from 10^{-4} to 10^{7} m/ka. This huge range of values reflects the increasing number and duration of intervals of non-deposition or erosion factored into the measurements as the length of the measured stratigraphic record increases. The sedimentary time scale, from 10^{-6} to 10^{6} years, constitutes a natural hierarchy corresponding to the natural time range of temporal processes—the burst-sweep cycle in turbulent transport, diurnal, lunar, seasonal, and geomorphic-threshold processes, orbital forcing, tectonism, etc. (Miall 1991a).

An increasingly rich and detailed record has become available in recent years pertaining to well documented recent and ancient deposits from which rates of stratigraphic and sedimentary processes can be extracted. It can now be demonstrated that the distribution of layer thicknesses, the durations of stratigraphic gaps, and sedimentation rates in stratigraphic successions accord to a fractal model (Miall, in press). This model provides an elegant basis for integrating the processes of accommodation generation with the data on varying sedimentation rates, the scales of hiatuses, and the processes that operate over these time scales (Fig. 2.7). The concept of the *Sedimentation Rate Scale* (*SRS*) has been proposed (Miall, in press) to encapsulate the rates and time scales over which sedimentological and stratigraphic processes occur (Table 2.1, Fig. 2.7), from the migration of grains across a bedform in a river channel (*SRS-1*) to the accumulation of basin-fill complexes (*SRS-10, 11*).

Most stratigraphic units, when accurately dated at geological time scales (10^{6}–10^{7} years) yield long-term, average sedimentation rates of 0.01 to 0.1 m/ka (10^{-1} to 10^{-2} m/ka: *SRS-11*). In some rapidly subsiding basins, such as the proximal regions of some foreland basins, pull-apart basins in strike slip settings, and a few other scenarios, long-term accumulation rates of > 1 m/ka (10^{0} m/ka: *SRS-10*) have been measured.

Throughout this discussion, sedimentation rates are presented as logarithmic values, both to emphasize the order-of-magnitude variations under discussion, and also to underscore the consistency of the rates defined for each *SRS* category.

The LAB models are based on *SRS-7*, that is, a time scale of $10^{3}-10^{4}$ years, and sedimentation rates of 10^{0}–10^{1} m/ka, characteristic of long-term geomorphic processes, including channel aggradation and avulsion, and the development and switching of channel belts. Studies of post-glacial alluvial systems fit within this time- and rate-scale definition, e.g., the Cumberland Marshes studies of Morozova and Smith (1999), and the Rhine-Meuse studies of Stouthamer and Berendsen (2007) and Stouthamer et al. (2011). The problem, then, is that the modern studies and the simulations on which the LAB and WMSM models are based assume an accommodation rate up to three orders of magnitude more rapid than is typically represented in the preserved ancient record.

A variation on the LAB models was presented by Marriott (1999), in which she modeled cyclic variations in base level at a sedimentation rate of 0.21 m/ka (10^{-1} m/ka). Four complete cycles of base-level change were modeled within a simulated time span of 474,650 years, and an avulsion about every 2400 years. This is within the time and rate scale of "4th-order" Milankovitch cycles, as noted by the author, and corresponds to *SRS-8*. However, this is still an order of magnitude more rapid than the case studies discussed below.

There is also the larger question of preservability. How much of the succession now being formed will survive glacial-interglacial cycles into the geological record? The accommodation that permitted modern coastal fluvial, estuarine and deltaic systems to aggrade was created by post-glacial sea-level rise, at a rate in the range of 5 m/ka (concordant with *SRS-7*). Tectonic subsidence, at a rate of 10^{-1} to 10^{-2} m/ka, continuing for a full-glacial cycle of 10^5 years, would provide, at most, a few metres to tens of metres of accommodation. But a full-scale glacioeustatic fall could remove most of the deposits by erosion, leaving only the oldest portion of the record, that corresponding to the lowstand or early transgressive phase. Over a geological time scale, successive similar fragments would become amalgamated between successive regional unconformities.

6.2.3 Some Examples from the Geological Record

Castlegate Sandstone, *Book Cliffs*, *Utah:* The Campanian portion of the Sevier clastic wedge, in the classic area of the Book Cliffs (Utah-Colorado) has long been interpreted as the product of repeated thrust loading along the Sevier fold-thrust belt (Yoshida et al. 1996; Horton et al. 2004). Aschoff and Steel (2011) calculated rates of coastal progradation and rates of sediment accumulation. The central portion of the section, which they term Wedge B, and which includes most of the Castlegate Sandstone, yields an accumulation rate 47 m/my (= 0.047 m/ka: *SRS 9, 11*) over 1.92 my.

Robinson and Slingerland (1998, Fig. 3) defined the stratigraphy slightly differently. Their Castlegate sequence, comprising 200 m of section at the type section (Price Canyon) was deposited in 4 my, yielding an accumulation rate of 0.05 m/ka (10^{-2} m/ka; *SRS 9*, *11*).

Olsen et al. (1995) subdivided the Campanian to Paleocene strata of the Book Cliffs into five sequences, of which the Castlegate Sandstone comprises one. They defined and subdivided the sequences based on a contrast between sandstone-dominated successions at the top and base of the sequences, and shale-rich middle portions, in which some evidence of tidal influence is present, in the form of tidal bedding and *Skolithos* traces (see also Yoshida 2000). The gradation between these contrasting facies assemblages was attributed to increasing and decreasing rates of base-level rise (Olsen et al. 1995, Fig. 10). Their model is illustrated here as Fig. 6.5.

Olsen et al. (1995, p. 276) state that "A similar model with a similar structure and dominance of the transgressive aspects, with more weight to the positions and

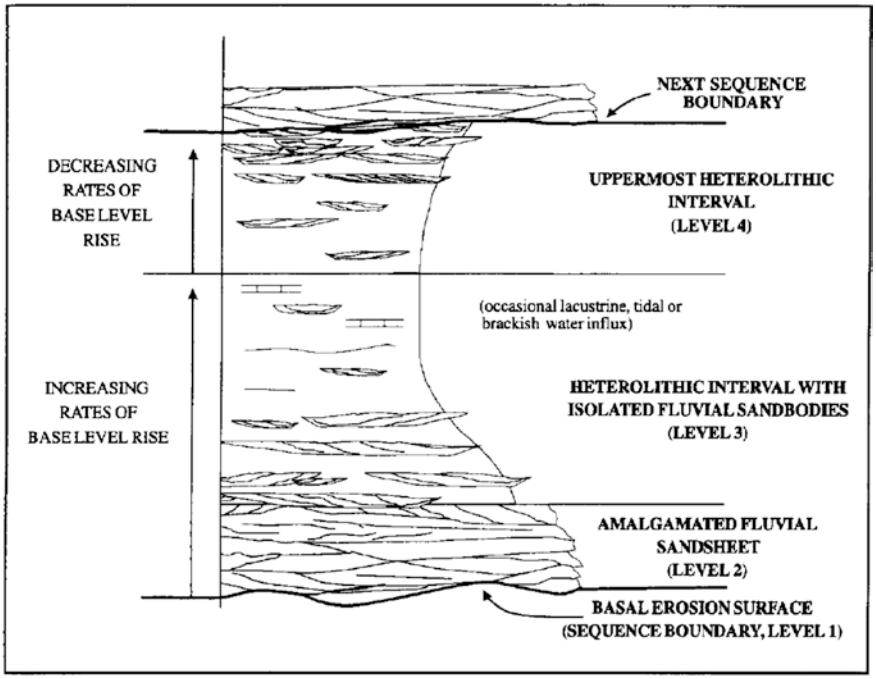


Fig. 6.5 Idealized fluvial sequence model for the Mesaverde Group, Utah (Olsen et al. 1995, Fig. 10, p. 276)

maturity of soils within a fluvial sequence, has been developed by Wright and Marriott (1993).” However, the sedimentation rates calculated above cannot be reconciled with this type of changing accommodation model. Avulsion periodicity (channel return rate) in modern systems is around 10^4 years and does not vary by orders of magnitude from this range in natural systems. Therefore, at *SRS 9* and *11*, channels will always return to a former position before the earlier deposit is buried, and therefore amalgamated architectures are always to be expected.

The sedimentological observations of Olsen et al. (1995, Fig. 10) and Yoshida (2000) require a different interpretation. Whereas a model based on changing rates of accommodation will not work, one base on shifting facies belts does. Sea-level change or tectonic adjustments to accommodation may cause influx and retreat of marine influence. Intraplate stress changes can generate accommodation at 0.01 to 0.1 m/ka at time scales of 10^6 yrs (Cloetingh 1988, p. 216), which is within the range of *SRS 9-11*. Episodic thrust loading within a foreland-basin setting may generate regional basement adjustments at a higher rate (Peper et al. 1992). This was the preferred model of Yoshida et al. (1996) and Miall and Arush (2001a) for the sequence architecture of the Castlegate Sandstone, and would complement the regional model of Aschoff and Steel (2011).

Morrison and Cedar Mountain Formations, *Utah-Colorado:* A sequence analysis of the Upper Jurassic and Lower Cretaceous strata of the Sevier foreland basin in the Uinta Mountains of Utah and Colorado was the basis from which Currie (1997) suggested the terms aggradational, transitional and degradational for nonmarine systems tracts, as alternatives to the standard marine terminology (Fig. 6.6). Currie (1997) defined four sequences; two are discussed here.

Sequence UJ-2 comprises the Upper Salt Wash and Brushy Basin members of the Morrison Formation. It was deposited during the Kimmeridgian-Berriasian, over a time span of approximately 10 my. The sequence ranges from 35 m in thickness in central Wyoming to 125 m in the area of the San Rafael Swell. These values indicate sedimentation rates ranging from 0.0035 to 0.0125 m/ka (10^{-3}

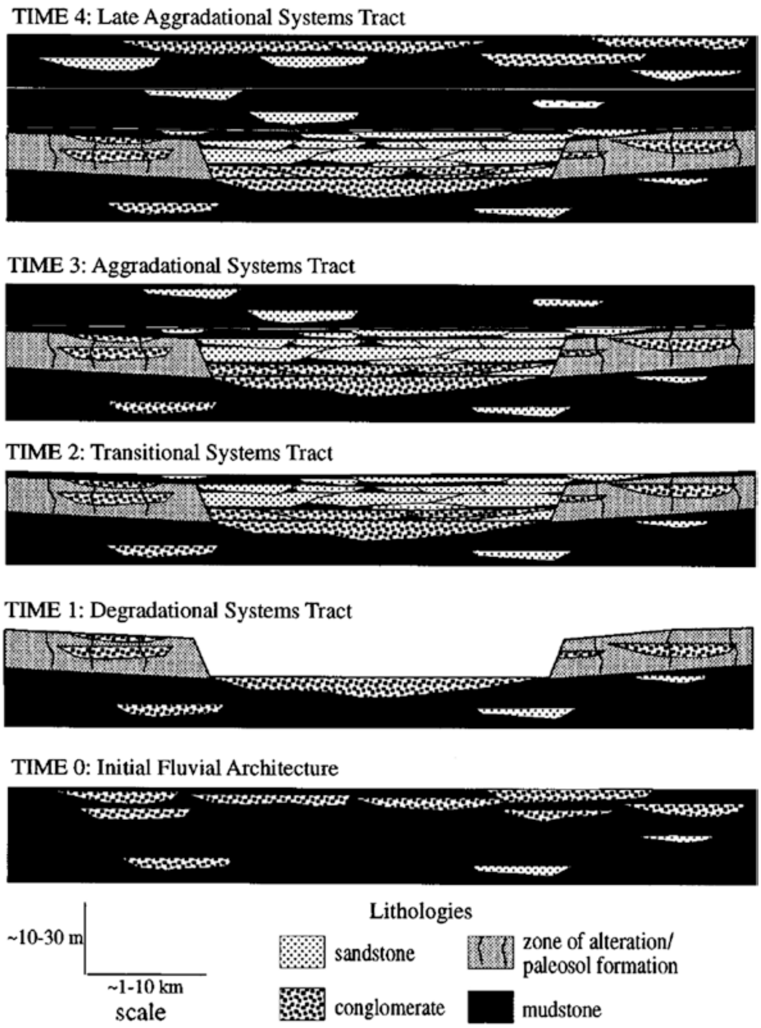


Fig. 6.6 An alternative systems-tract terminology for fluvial systems (Currie 1997, Fig. 2, p. 1208)

to 10^{-2} m/ka). Sequence LK-2 corresponds to the Cedar Mountain Formation, together with the Cloverly Conglomerate in the east. It ranges in thickness from 30–90 m within the project area and was deposited over a time span of 16 my, indicating a sedimentation rate of 0.0019 to 0.0056 m/ka. Beyond the project area, at the west end of the Uinta Mountains, isopachs indicate a thickness exceeding 300 m, which yields a sedimentation rate of 0.03 m/ka (10^{-2} m/ka). These values are those of *SRS 9*, *11*. Unconformities representing nearly 10 my are present between sequences UJ-2 and LK-1 and between LK-1 and LK-2. Accommodation generation was therefore slow to negative in the project area throughout the Upper Jurassic to Lower Cretaceous.

Currie (1997) interpreted the deposits as the product of tectonically-controlled sedimentation over the forebulge and backbulge of the Sevier foreland basin. They state (p. 1207): "high rates of aggradation will produce channel sandstones that are isolated both vertically and laterally by fine-grained overbank material (e.g., Allen 1978; Bridge and Leeder 1979)." On p. 1208 they further state: "The assumption behind these models is that channel avulsion frequency is random and independent of sedimentation rate. However, increased sedimentation rates associated with an accommodation increase may promote more rapid rates of channel avulsion and a higher proportion of channel interconnectedness (Heller and Paola 1996)." They concluded (p. 1212) that "four depositional sequences [are] defined by changes in sedimentary architecture that resulted from variations in the rate of accommodation development."

However, the facies assemblages defined as aggradational, transitional and degradational cannot be interpreted in terms of the LAB model. Much slower processes were at work. It seems more likely that the facies variation reflects long-term (10^6-year) tectonic controls on paleoslopes, sediment supply, and paleogeography. This is clearly implied by Currie's (1997) analysis. The comparisons to the LAB model can therefore be considered an inappropriate distraction.

Ericson Sandstone, *Rock Springs Uplift*, *Wyoming:* This fluvial to delta-plain unit was subdivided into two sequences by Martinsen et al. (1999), and interpreted in terms of cyclic fluctuations in the accommodation/supply ratio in response to changes in the rate of base level rise and fall (Fig. 6.7). They stated (p. 248) "We suggest that in non-marine successions, the sequence architecture can be divided into low- or high-accommodation systems tracts based on alluvial architecture."

Estuarine and deltaic deposits, including single-storey sandstones were interpreted as high accommodation deposits, and fluvial sheet sandstones representing amalgamated channel sandstones were interpreted as low accommodation deposits. Explanations of these variations were made with reference to the LAB models.

The Ericson Formation and the lower Almond Formation (which is included in the sequence analysis) span much of the upper Campanian. Their stratigraphic table indicates that the deposits span 15 of the 21 ammonite zones which comprise the Campanian. Therefore the deposits represent approximately 9.2 my. The formation ranges from 100 to 400 m in thickness within their project area, which indicates average sedimentation rates of 0.011 to 0.043 m/ka. The section includes a major disconformity (sequence boundary) which may represent as much as half

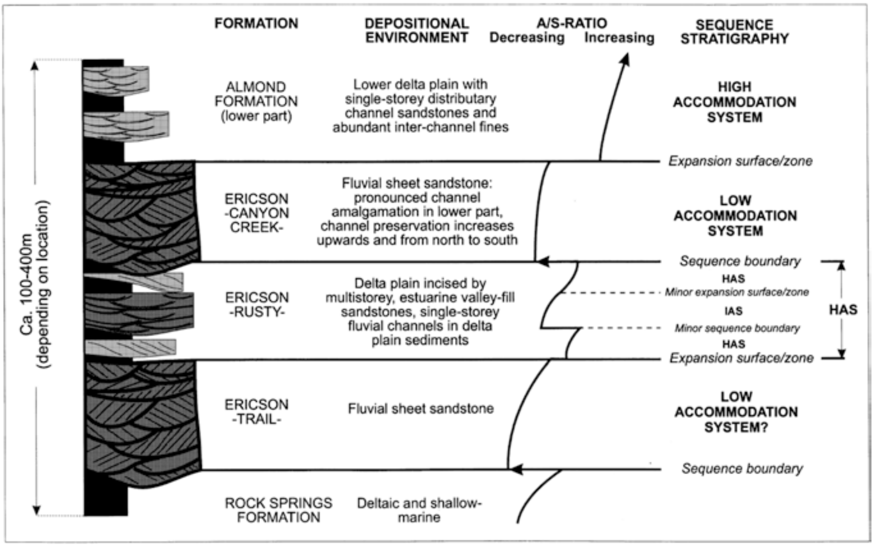


Fig. 6.7 Sequence stratigraphic model of the Ericson Sandstone in the Rock Springs Uplift, Wyoming (Martinesen et al. 1999, Fig. 14, p. 255)

of the elapsed time, indicating that sedimentation rates for the preserved portion of the record could have been double the figures quoted above. But this is still within the range of 10^{-2} m/ka and corresponds to the *SRS 9* or *11* time scale.

Based on the arguments presented here, interpretation of the changes in facies assemblage based on changes in the accommodation/supply ratio may be challenged. However, in arguing for a regional control by repeated tectonic activity within the nearby Wind River Range the authors are probably correct. As with the previous case, paleoslopes, sediment supply, and paleogeography may be controlled on a 10^6-year time scale.

Williams Fork Formation (*Upper Cretaceous*), *Piceance basin*, *Colorado:* Subsurface data revealed a pattern of channel clustering and compensational stacking in this formation (Fig. 3.27), which Hofmann et al. (2011) attributed to changing rates of accommodation, with reference to the LAB models. However, the rate of change they suggested (~400 ka) does not correspond to that of the LAB models and they suggested allogenic, rather than autogenic, mechanisms underlying the generation of these architectural patterns.

Permo-Triassic, *Iberian Basin*, *Spain:* López-Gómez et al. (2010) described the fluvial fill of a Permian–Triassic basin in eastern Spain that subsided at varying rates as a result of episodic extensional faulting and thermal subsidence. The most rapid subsidence rates, in the range of 0.05 m/ka, occurred over a 2-my period in the late Permian (256–254 Ma; López-Gómez et al. 2010, Fig. 6). Subsidence rates during the Triassic were at about half this rate, but still, at 10^{-2} m/ka, within the *SRS 11* range. This study explored the relationship of fluvial architecture to stretching factors, using values of δ and β for upper- and lower-crustal stretching factors calculated from borehole data, it being assumed that the architecture would be a proxy indicator of subsidence rates. However, subsidence rates were neither quoted nor discussed in the paper (those cited here were calculated from data presented in their Fig. 6), and the variations in alluvial architecture (Fig. 2.38 of this book) do not seem to this writer to suggest any systematic relationship to the stratigraphy.

The authors suggested a relationship between architecture and stretching factor, but the relationships do not appear to be straightforward. Indeed, they noted that other factors, including climate change, could have affected the final composition of the preserved alluvial architecture. Subsidence rates varied within one order of magnitude, values at least an order of magnitude smaller than the *SRS-7* values required to generate and preserve architectural differences using the LAB and standard sequence models.

Iwaki Formation, *Eocene*, *northeast Japan:* A 400-m thick conglomeratic unit, of Middle-Late Eocene age, shows a distinct upward trend from laterally-amalgamated, multistory channels, to multistory channels, to isolated, single channels (Komatsubara 2004). No precise age information was provided, but if the assignment of a mid-Late Eocene age is used, the sedimentation rate was approximately 0.06 m/ka (10^{-2} m/ka). The author invoked the WMSM sequence models but, as with the previous cases, we can now argue that this comparison is inappropriate.

Ishikari Group, *Eocene*, *Hokkaido*, *Japan:* The Ishikari Group consists of a complex of estuarine to fluvial deposits, and was subdivided into four sequences by Takano and Waseda (2003). The youngest of these sequences (lsk-4), totaling as much as 1450 m, was deposited between about 39.5 and 37 Ma, indicating a sedimentation rate of 0.58 m/ka (10^{-1} m/ka). This is a geologically rapid rate, as noted by the authors, but is consistent with a geological setting comprising a foreland basin that developed over a forearc following arc-continent collision.

A sequence model for these estuarine to fluvial deposits was proposed (Fig. 6.8). Changes in fluvial style and channel density through the uppermost sequence (Takano and Waseda 2003, Fig. 14) were compared by the authors to the LAB models and the WMSM sequence models. The rapid rate of subsidence and the time frame are consistent with the high rates of convergent-margin basins, and constitute a good field example *SRS 10*. The sedimentation rate is comparable to that of *SRS 8* and *9*. However, only one complete architectural cycle is represented by sequence lsk-4, which spans 2.5 my, compared to the orbital cycles of *SRS 8* and *9* and the 10^4-yr autogenic cycles of the LAB models. The comparison is therefore not appropriate.

Camp Rice Formation (*Upper Pliocene-Lower Pleistocene*, *Rio Grande Rift*, *New Mexico*: Stratigraphic events in this rift basin are well constrained by magnetostratigraphy. Mack and Madoff (2005) analyzed channel distribution and paleosoil maturity and tested the data against the fluvial architecture models of Bridge and Leeder (1979) and Bryant et al. (1995). Sedimentation rates over a study interval of 50 m representing approximately 1.6 my ranged from 0.036 m/ka at the base of the section to 0.017 m/ka at the top. This study therefore represents

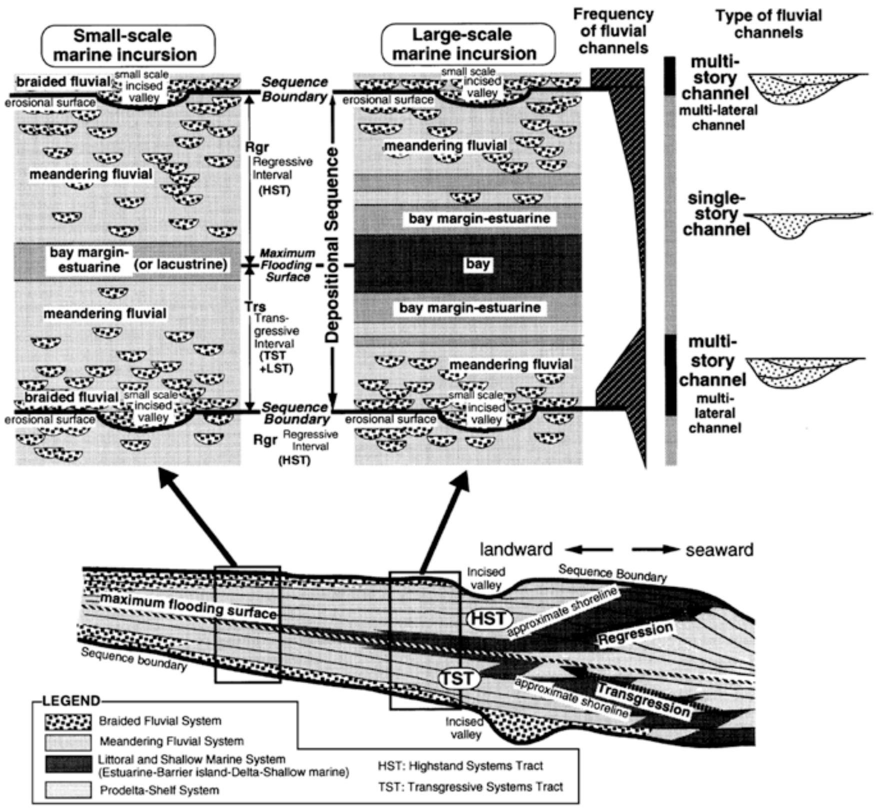


Fig. 6.8 Sequence model for the Ishikari Group (Takano and Waseda 2003, Fig. 13, p. 152)

a focus at the *SRS-9* scale. The authors stated (p. 207) that their study provided "a test of numerical, theoretical, and experimental models of fluvial architecture and palaeosol development." However, the time scales represented by this basin fill are an order of magnitude greater than those of the LAB model. Channel return rates ranged from 228–685 ka in different sections measured in this rift valley succession. Channel sand bodies are separated by mature paleosoils indicating periods of landscape stability lasting 10^4–10^5 years. These values suggest that what is being measured is allogenic shifting of channel belts, not individual channels, in response to tectonic tilting of the basin.

Eocene of Tremp-Ager Basin, *South Pyrenees*, *Spain*: Nijman (1998) developed a detailed relationship between tectonism, the evolving paleogeography, and the resulting stratigraphy for the Eocene fill of the Tremp-Ager Basin in Spain. In this case, long-term geological rates and shorter-term sedimentological rates appear to be comparable. The overall sedimentation rate for the 2000 m of section, which accumulated in 10 my is 0.2 m/ka = 10^{-1} m/ka (Nijman 1998, Fig. 13). The time and the rate are equivalent to *SRS 9* or *10*. Individual cycles in this succession

(Nijman 1998, Table 2) lasted between 100 and 150 ka at sedimentation rates of 0.185–0.673 m/ka = 10^{-1} m/ka, equivalent to *SRS-8*, the time and rate scale of which encompass high-frequency orbital cyclicity. Nijman (1998) described complex variations in alluvial architecture, including aggradational and amalgamated sheet cycles, and attributes the changing architecture primarily to tectonic influences. The LAB and WMSM models were not invoked in this case.

Plio-Pleistocene of the Nyírség-Pannonian Basin, *Hungary*: Closely-spaced groundwater wells and good wireline log control enabled Püspöki et al. (2013) to develop a detailed chronostratigraphy for the Plio-Pleistocene section on the northeastern margin of the Pannonian Basin in Hungary. Figure 6.9 is a strike-oriented cross-section from near the margin to near the centre of the Nyírség sub-basin. Facies analysis suggests a subdivision into two types of systems tract deposits, one developed in a low-accommodation setting and the other in a high-accommodation setting. Low accommodation deposits rest on sequence boundaries and consist of amalgamated channel deposits. High-accommodation deposits are characterized by complete macroform (point-bar) deposits, heterolithic complexes, and a greater preservation of overbank fines. Near the centre of the section one unit of high-accommodation deposits is interpreted as lacustrine in origin. The sequence boundaries at the base of the low-accommodation successions are erosional, and have stepped margins, as seen in Fig. 6.9. These are

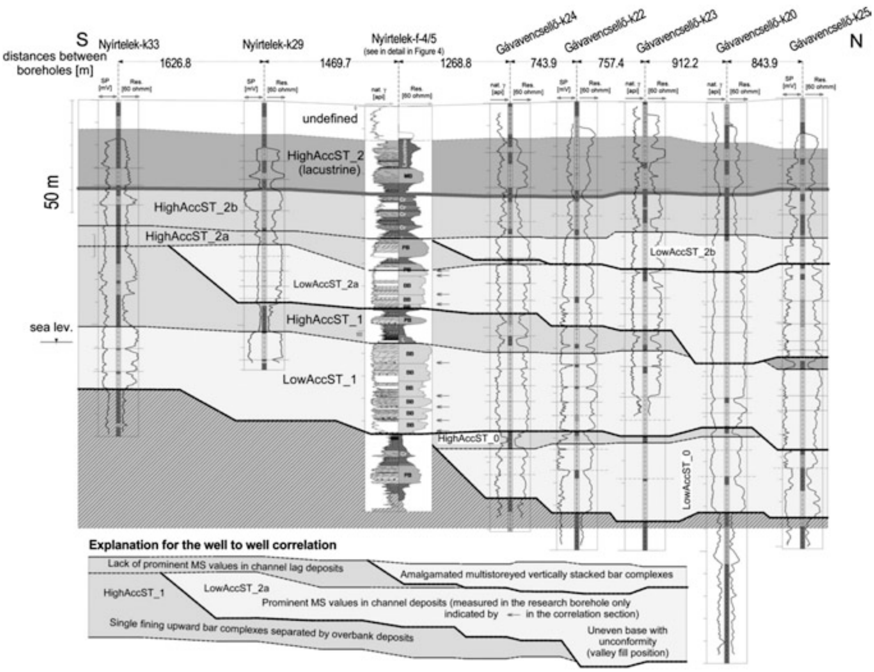


Fig. 6.9 (Püspöki et al. 2013, Fig. 7, p. 41)

interpreted as erosional terraces, the elevated terraced margins corresponding to interfluve areas, on which mature paleosoils were recorded in some boreholes. Based on magnetic susceptibility measurements a regional correlation has been established with the Chinese loess succession (Fig. 6.10), and this, in turn, has been correlated to the global oxygen isotope scale, which provides an absolute chronology for the section (Fig. 6.11). This has demonstrated that the Hungarian basin succession is almost complete, in that virtually all the chronostratigraphic stages have been recognized. Sequences ranged in duration from 120,000 to 730,000 years. The loess chronology is based on a much higher-frequency Milanokovitch signal, and it is clear from Fig. 6.11 that the sequence generating mechanism is not orbital forcing—the timing is wrong, although, as noted below, it seems likely that climate change may have been an important secondary control on sequence generation. Rates of sedimentation are in the range of 0.17 m/ka (10^{-1} m/ka), (based on the ages of the sequence boundaries and the

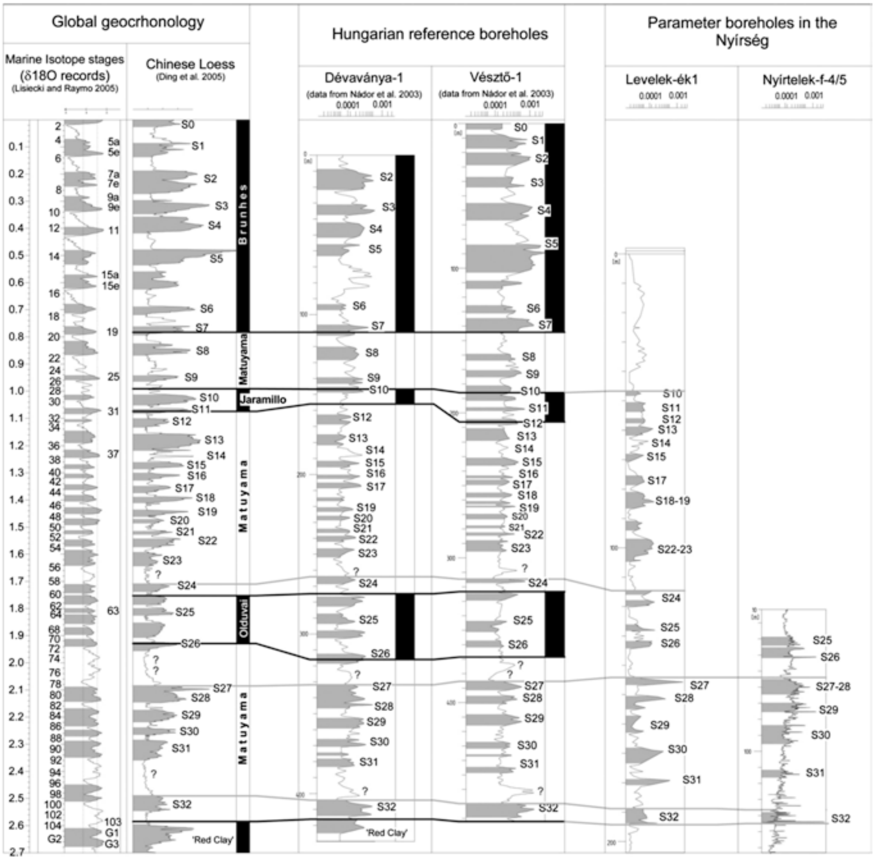


Fig. 6.10 (Püspöki et al. 2013, Fig. 14, p. 48)

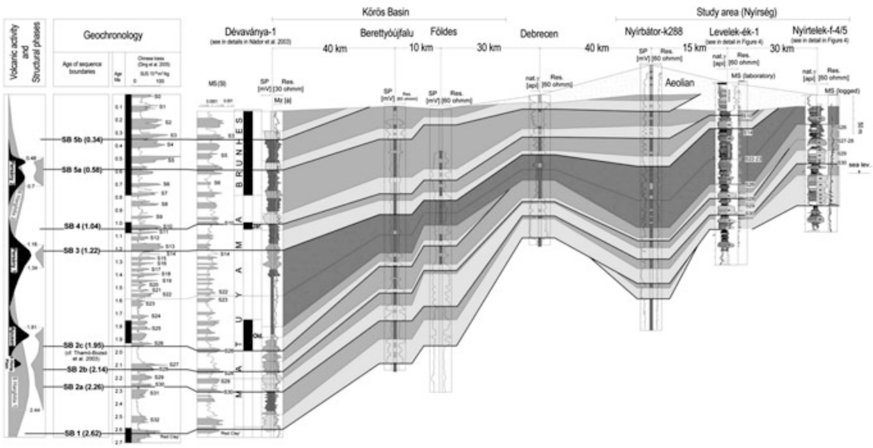


Fig. 6.11 (Püspöki et al. 2013, Fig. 15, p. 49)

thicknesses of the sequences) which corresponds to *SRS 9-10*, equivalent to the rates measured in tectonically very active basins. Püspöki et al. (2013) interpreted the sequence-generating mechanism as tectonic in origin, the sequence boundaries being generated at times of active uplift, which in part correspond to times of volcanic activity. This is summarized in Fig. 6.12. The likely course of the major rivers can be reconstructed through time by mapping the distribution of the high- and low-accommodation facies. The "areas of frequent fluvial denudation" in the block diagrams are the interfluve areas prone to erosion and terrace formation.

Püspöki et al. (2013) cite earlier work that suggests an association of climate change from dry, cold episodes to warmer, humid times, with the generation of sequence boundaries. Although the sequences are not orbitally forced, they suggested (p. 52) that

> In dry and cold periods, due to the low transport capacity of the rivers, coarse grained sediments were trapped in the updip parts of the basin, close to the areas of sediment entry points of the supplying systems. This sediment was then transported into the down dip parts of the basin during the subsequent warm and humid periods.

Climate change may have acted as a trigger, that pushed the rate of landscape change to a point where it reached a threshold that accelerated the much slower rate of change driven by tectonism (Schumm 1979). The evidence for this assertion is that the sequence boundaries appear to correlate to the base of intervals of high magnetic susceptibility. These high MS values are, in turn, correlated to interglacials on the oxygen isotope scale.

This interpretation does not rest on an assumed association between autogenic channel stacking pattern and accommodation generation (the basis of the LAB models), and is in accordance with the interpretations of alluvial sequence architecture developed in this chapter.

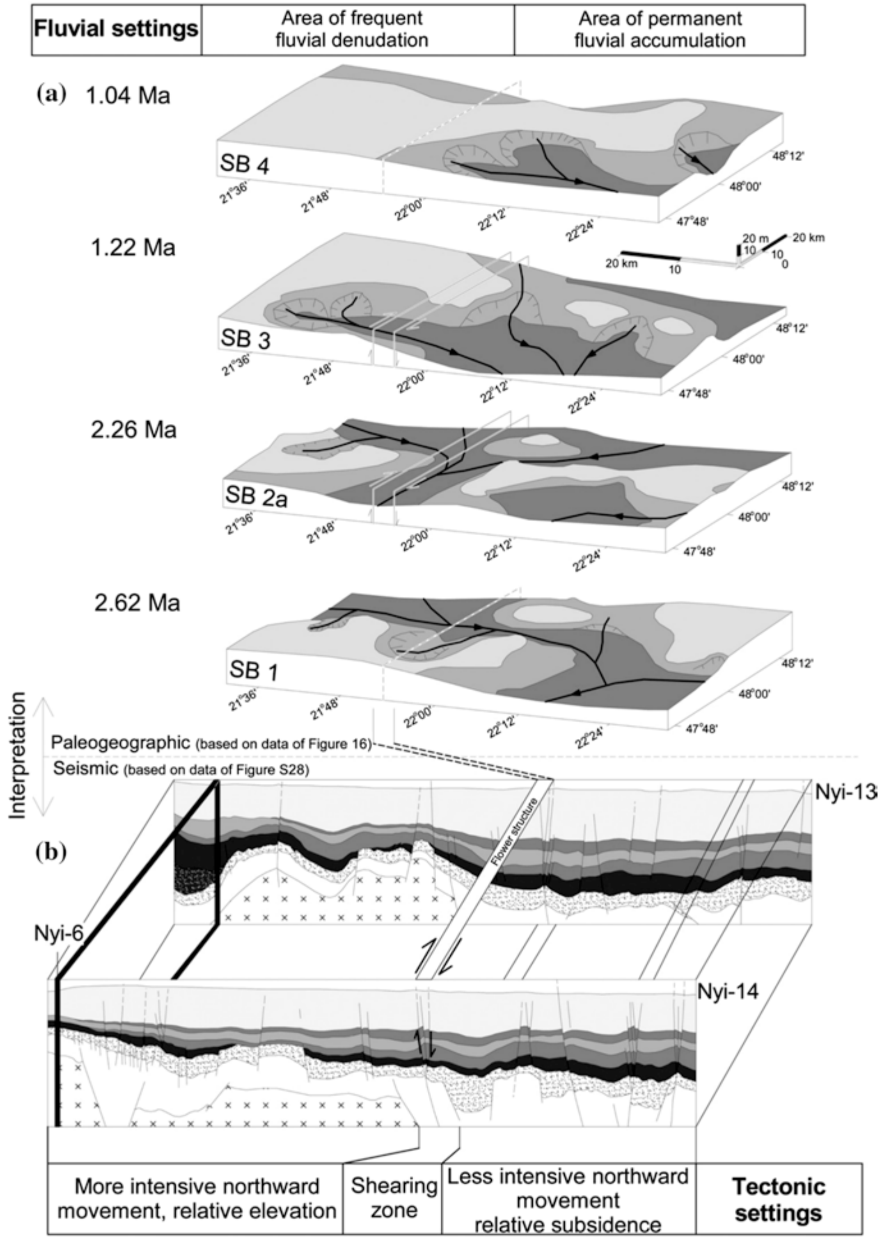


Fig. 6.12 (Püspöki et al. 2013, Fig. 17, p. 51)

6.2.4 Discussion

In a significant way the present is not the key to the past. There is an orders-of-magnitude difference in sedimentation rates between those measured from modern sediments and the post-glacial record, and those recorded from the ancient geological record. The high-frequency processes that form the basis for the LAB models, based in turn on the post-glacial record, are not preserved from the more remote geological past. High accumulation rates occur regionally in only a very few geological settings, such as in the source-proximal corners of some convergent-margin basins, and it requires special circumstances for the shorter-term high frequency processes that constitute *SRS 1* to *7* to achieve long-term preservation (Miall, in press).

Gibling (2006, p. 761) made a related point. Focusing on the many controls on alluvial architecture, including discharge, sediment supply, bank materials, and so on, he warned that: "The recent tendency in sequence stratigraphy to relate channel-body form to accommodation (e.g., Shanley and McCabe 1994) is thus subject to many caveats." There is not necessarily a simple relationship between forcing processes and a stratigraphic result. Jerolmack and Paola (2010) refer to the "Shredding of environmental signals by sediment transport" (the title of their paper). Gibling et al. (2011, p. 439–441) offered similar comments, suggesting that the alluvial stacking pattern on which accommodation models have been based may in fact be controlled by climate change or changes in fluvial paleogeography unrelated to subsidence rates. They suggested that the Quaternary record, characterized by an icehouse climate, with high-magnitude and high-frequency environmental change, is not necessarily a good model for interpreting the ancient record. Valleys formed by Neogene glacial sea-level cycles are likely to be a more prominent part of the coastal sedimentary record that during the Mesozoic, for example, when it assumed that glacioeustasy was of minor importance to non-existent.

It is argued here that systematic stratigraphic changes in alluvial architecture in the ancient rock record are not the product of changing avulsion rates and changes in fluvial style under the influence of variable rates of accommodation, as in the LAB models, but reflect regional shifts in facies belts, that themselves are a response to tectonism and to changes in accommodation and other variables (climate, discharge, sediment supply, etc.) at much slower rates, by one to several orders of magnitude, than those assumed for the LAB models. A model of a more realistic interpretation of the reasons for regional changes in fluvial style through time is provided by the interpretation of shifting facies belts of the Loranca fluvial system (Fig. 3.13). This system was most likely controlled by tectonism.

Some recent studies of ancient fluvial systems are beginning to provide the essential sedimentological data to support the argument made in this section. For example, Hampson et al. (2013) documented in detail the changes in fluvial style through the Blackhawk Formation of central Utah. This unit is part of the Mesaverde Group (as is the overlying Castlegate Sandstone, discussed earlier in this section), a very well documented succession for which a detailed sequence-stratigraphic framework is available, from which may be derived testable hypotheses about the relationship between fluvial style and accommodation. Outcrop quality enabled some detailed reconstructions of alluvial architecture to be carried out, and these revealed a range of fluvial styles throughout the Blackhawk Formation, but no systematic changes with stratigraphic position. They also made reference to earlier studies (Adams and Bhattacharya 2005; McLaurin and Steel 2007) which demonstrated no systematic variation in fluvial style through the overlying sequence boundary into the Castlegate Formation. The conclusion of Hampson et al. (2013, p. 166) is that the "internal architectures of the sandbodies do not result from systematic, short-term changes in accommodation such as those that characterize incised-valley fills formed by relative sea-level change in coastal-plain settings" but they they "appear to reflect local changes in the balance of sediment flux and transport capacity due to upstream controls, such as high-frequency climatic variations, and autogenic responses."

One of the experimental stratigraphy studies by Paola's group has helped to confirm this interpretation. Strong et al. (2005) ran an experiment to construct a braided fluvial fan delta under varying subsidence rates and different conditions of water and sediment discharge. The experimental sediment supply consisted of a mixture of two lithologies, quartz sand, and coal particles, the latter serving as a proxy for fine-grained sediment by virtue of its lower density. Four different sets of background conditions were used in the experiment, the results of which were then sliced vertically, parallel and perpendicular to depositional strike, to construct the cross-sections shown in Fig. 6.13. The most interesting outcome of this series of experiments is the transition from Stage 2 to Stage 3, the cause of which is discussed below.

The enlarged cross section of Stage 2 in Fig. 6.13 illustrates a simulated alluvial plain with stacked channel deposits, highlighted in black to increase their visibility (in other words, these are sand channels, not coal). Stage 3 deposits consist primarily of laminated coal layers, with no visible channels. The section parallel to depositional dip (the bottom panel in Fig. 6.13) shows a major lateral shift in the facies boundary. Stages 1 and 2 deposits are primarily sand, whereas Stage 3 deposits are coal-rich, with a transition upstream into sand taking place half way up the experimental delta. These architectures would appear to illustrate the contrasting conditions in the LAB models of stacked channels (Stage 2) indicating slow subsidence and channel amalgamation, versus rare, dispersed channels (Stage 3) corresponding to the condition of rapid subsidence. In fact, the change in conditions from Stage 2 to Stage 3 was exactly the opposite. This transition represents a reduction of about 75 % in the subsidence rate, together with a reduction in the discharge and sediment load. With reduced slope and water in Stage 3, most of the sand was deposited in the proximal part of the delta, with the coal portion selected out and transported further downslope. The transition to Stage 3 represents an upstream shift in the facies boundary between sand-rich, channelized deposits and distal, fine grained deposits with few to no channels. The relevance to full-scale alluvial systems is that it is the overall balance of conditions that determines alluvial architecture, subsidence rate being only one of them.

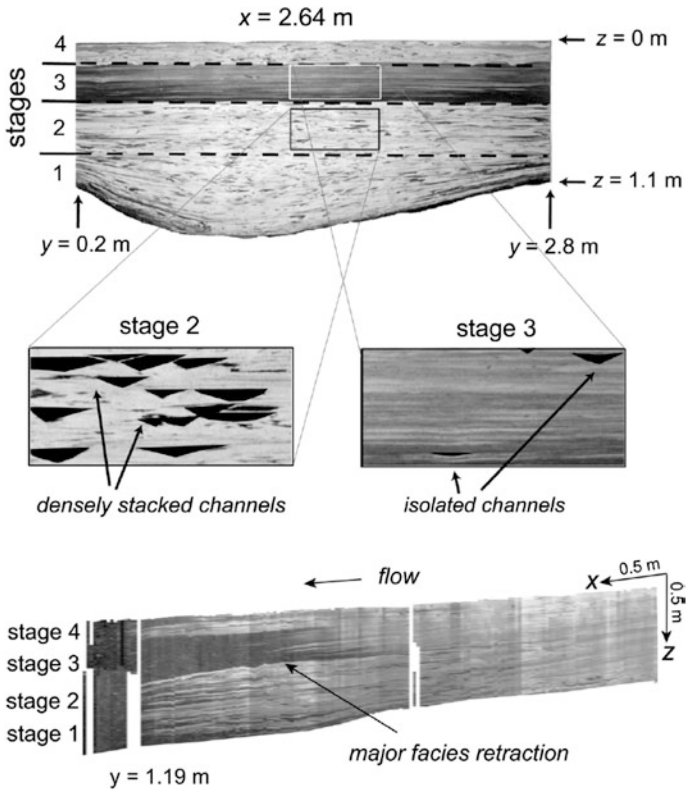


Fig. 6.13 The results of an experiment to model variations in alluvial architecture base on variations in subsidence rate, water discharge and sediment load. The experiment was run under four successive combinations of conditions. The *top three diagrams* illustrate a strike-parallel section through the experimental result, the *bottom diagram* is a dip-parallel section (Strong et al. 2005, Figs. 3, 4, p. 247)

6.3 Sequence Stratigraphy of Ancient Fluvial Systems

6.3.1 How to Re-Focus Fluvial Studies of the Ancient Record

Given the discussion of the previous sections of this chapter, it should now be clear that there are a number of hitherto unrecognized difficulties in studying sedimentation in the distant past. We need to take on board several important points with regard to the interpretation of the ancient record:

- Although the concept of "the present is the key to the past" can be used with confidence to interpret the products of basic processes such as those related to the flow of water, turbulence, bedform generation, soil generation, and so on, all of which are built on standard principles of physics and chemistry, we need

to be much more cognisant than we have been to date of the importance of time and of the rates of processes.
- The fragmentary nature of the preserved record means that we need to be much more cautious than we have been with regard to assumptions about continuous sedimentation and the interpretations that include this as a starting point, such as the cyclicity that may be observed in vertical sections and which has become an essential component of fluvial facies analysis methods.
- Fluvial environments and processes are likely to be highly variable across a basin at a one moment in time, and also to vary significantly through time. As Gibling et al. (2011, p. 441) noted: "The identification of systems tracts in an alluvial basin tends to imply a basinwide effect, whether the authors intended this or not. However, alluvial regions such as the Ganga Plains are huge areas with many geomorphic elements (orogenic and cratonic sources; megafans, interfluves and axial drainage systems) and large precipitation gradients. In basins of this scale, basinwide bounding surfaces will be a rarity, and most prominent surfaces may be of local extent."
- The high rates of processes recorded in the post-glacial record, and the high-frequency and high-amplitude changes in climate and sea-level that characterized the Neogene, are largely non-uniformitarian, in that most of geologic time was probably not characterized by such large-scale changes. With the exception of the Gondwana glaciation of the Carboniferous to Permian, most of the Earth's continents throughout most of the Phanerozoic appear to have been characterized by a greenhouse climate (exceptions to this are discussed below). This means that we must be very careful in the way we use Late Cenozoic analogs as a basis for interpreting the ancient record.
- As has now been very well documented by Davies and Gibling (2010a, b, 2011), the evolution of land plants had a profound effect on the shaping of the alluvial landscape, with major changes occurring through the mid-Paleozoic.

Allogenic and autogenic sedimentary processes may generate predictable, ordered stratigraphic patterns at all time scales. The order and predictability may include erosional processes as well as processes of accumulation. This has always been the basis for Walther's Law and, more recently, sequence stratigraphy. Therefore, contrary to the random or chaotic processes of accumulation implied by Bailey and Smith (2010), stratigraphic order, including cyclicity, may be preserved in the rock record and may be understood and interpreted within the focus of the appropriate *SRS*. Walther's Law is based on the concept of shifting depositional environments represented by deposits stacked in a vertical succession. Many of the environments to which this law has been applied generate the characteristic vertical profiles (e.g., fluvial fining-upward cycles, deltaic crevasse splays) over time spans of 10^1–10^3 years, and may, therefore be interpreted within the framework of *SRS 5-7.* Larger scale systems, such as deltas, develop in accordance with Walther's Law at longer terms, *SRS 7-9*.

The greenhouse climates of the Phanerozoic may not have been greenhouse all the time. Miller et al. (2005a, b) argued for a new interpretation of Cretaceous to early Cenozoic climates, based on their detailed studies of New Jersey stratigraphy, including the oxygen isotope record. While acknowledging the sparseness of the isotopic data, and the susceptibility to diagenetic modification of samples from older, relatively deeply buried sections, they argued that enough evidence had now been assembled for a rethinking of the concept of a long period of greenhouse climatic conditions. They suggested that the climate was characterized by "cool snaps"—the development of small ice caps lasting in the order of 100 ka. It was postulated that these ice caps must have been located in Antarctica, which was, as it is now, located in a polar position. The implication is that cycles of glacioeustatic sea-level change might be expected for parts of the Mesozoic. Amplitudes would be expected to be modest—a few tens of metres, at most.

The detailed stratigraphic studies that have been carried out by A. G. Plint and his group on the Cretaceous record of northwestern Alberta have yielded many clues to allogenic processes operating in that area. In several of his studies (Plint 1991, 2002; Plint et al. 2001), the evidence has suggested eustatic sea-level changes as a significant controlling mechanism. In a detailed study of paleovalleys (referred to later in this chapter), Plint et al. (2001) suggested that "eustatic excursions of the order of 20–30 m might have effected the base level changes responsible for valley formation, and suggested that valley incision took place throughout the falling stage systems tract" (Plint 2002, p. 294). The lack of evidence for structural tilting within this active foreland basin and the fact that the valleys maintained comparable depths throughout their length suggested that flexural movement of the basin (which might have been expected to be significant in this foreland-basin setting) was not a factor.

Given the conclusions from the very detailed studies cited in this chapter, is seems likely that relatively small-scale glacioeustatic cycles may be one of the processes that need to be considered in interpretations of at least the Cretaceous record of coastal fluvial systems. Further, the now extensive record for orbital forcing of lacustrine and other systems for many ancient successions through the Phanerozoic (e.g., see Miall 2010b, Chap. 7) suggests that orbital forcing of fluvial cyclicity, including effects on fluvial discharge and sediment load, with or without glacioeustasy, is likely an ubiquitous component of the driving processes involved at all times in fluvial sedimentation, although the effects were probably overprinted for much of geologic time by the more pronounced effects of local tectonism. The processes of "shredding of the environmental signal" referred to by Jerolmack and Paola (2010) also seems likely to have complicated the processes by which allogenic forcing is recorded in the preserved deposits. Allen (2008) demonstrated that the response time of alluvial systems to tectonic forcing may be significant, extending to at least 10^5 years, which means that high-frequency processes may be damped out completely.

Other allogenic mechanisms of relevant to the interpretation of the pre-Neogene record are discussed in Chap. 5, and include the effects of flexural loading and intraplate-stress changes in tectonically active basins. There is some evidence, discussed in that chapter, that the effects of high-frequency allogenic processes have been recorded in the fluvial sequence record, but this is clearly an area that needs considerable additional research.

6.3.2 The Sequence Boundary

Bounding surfaces are of critical importance in sequence stratigraphy, and there has been much debate regarding the appropriate surfaces to use in sequence definition, and how they relate to each other (Catuneanu 2006). In fluvial systems, the definition of sequences is, at least superficially, more straightforward. Regional subaerial erosion surfaces are the only bounding surfaces that have been identified in fluvial systems that have the necessary properties of being (1) laterally extensive and (2) apparently amenable to interpretation as the product of allogenic forcing, that they have been employed as essential components of the currently popular fluvial sequence models (e.g., Figs. 6.3, 6.4). Subaerial erosion surfaces are interpreted as representing episodes of negative accommodation. They are the surfaces exposed by falling base level or by uplift of the alluvial valley, such that the buffer zone of the alluvial system (in the terminology of Holbrook et al. (2006)) is perturbed downwards, in favour of erosion. The topography of the subaerial erosion surface reflects the pattern of erosion that develops during this negative accommodation phase, and the nature of the deposits immediately above and below the surface may provide essential information regarding the fluvial style, drainage patterns and the climate of the basin during the erosional interval.

It is common practice to regard sequence boundaries as chronostratigraphic surfaces for the purpose of mapping, but although this is a reasonable generalization for use on a broad, regional scale, it is important to make note of the important qualifications that modify this assumption. As Catuneanu (2006, p. 116) noted, "The stratigraphic hiatus associated with the subaerial unconformity is variable, due to differential fluvial incision and the gradual expansion of subaerial erosion in a basinward direction during the stage of base-level fall." As is the case for any unconformity, the age of the beds truncated by the unconformity may vary significantly from place to place, indicating a diachronism in the initiation of erosion at the surface, or variation in the depth of erosion. Similarly, the age of the beds that rest on the surface may also vary significantly from place to place, indicating a diachronism in the development of positive accommodation at the beginning of a sequence cycle; for example, a gradual transgression up an incised valley. As noted above, Gibling et al. (2011, p. 441) have pointed out the likely regional nature of key events and their sedimentological response in large alluvial basins.

If there is a lateral transition into marine deposits, the age of the beds below the sequence boundary and that of the first deposits above the sequence boundary is likely, in both cases, to be older at the coast than anywhere else. This is because a fall in sea level at the coast may initiate a wave of erosion that gradually translates inland up the river valley as a "knickpoint," resulting in longer and deeper erosion where erosion commences, at the coast (Fig. 5.34). Similarly, the bottoming out of a sea-level curve re-establishes a buffer zone at the coast, within which lowstand deposition can take place, and as transgression commences, this zone of sedimentation will gradually be extended up the valley inland.

Strong and Paola (2008) pointed out an important distinction between a *topographic surface* and a *stratigraphic surface*. The topographic surface that corresponds to the subaerial erosion surface undergoes continual change until it is finally buried and preserved. It is the stratigraphic surface that we map in the rock record, but this is a surface that never actually existed in its entirety as a topographic surface in its final preserved form, because it undergoes continuous modification by erosion or sedimentation until final burial (Fig. 6.14). The subaerial erosion surface may also violate one of the fundamental principles of an unconformity, which is that all the deposits below the unconformity surface are older than all the beds above the surface. Deepening and widening of an alluvial valley may continue during the final stages of the evolution of a subaerial surface, even while a turn-around in the base-level cycle has begun to transgress and bury the surface during the beginning of a transgression. Channel or overbank deposits that are preserved as terrace remnants, resting on the basal erosion surface, could therefore predate the coastal deposits formed during the final stages of base-level fall, and would therefore be older than the sequence boundary at the coast, although resting on it.

Many nonmarine sequence boundaries are clearly expressed as incised valleys, although where data are limited, it may be very difficult to distinguish an incised valley from a simple channel scour. This is discussed further, below. Given sufficient time (which could mean many millions of years) fluvial erosion may lead to peneplanation of the landscape, with the development of regionally significant, topographically flat unconformity surfaces. An erosional sequence boundary is therefore characterized by fluvial valleys that may range from narrow, shallow, isolated, linear valley architectures to broad, compound valleys, to undulating peneplaned surfaces (Sect. 6.3.3). Ancient examples are discussed at length in Sect. 6.3.4. Where erosion of the interfluve areas is slow, because the landscape has achieved the condition of peneplanation or because fluvial energy is low, reflecting low regional slopes, much of the surface may be capped by a soil. The study of paleosoils can yield much useful information regarding regional climate, as discussed in Sect. 5.2.2. Paleosoils may constitute useful markers of sequence boundaries, as discussed in Sect. 6.3.6. In other cases, sequence boundaries may be marked by major changes in fluvial style, with little or no evidence of channel incision. As discussed in Sect. 6.3.5, such non-erosional sequence boundaries, including what Miall and Arush (2001a, b) termed *cryptic sequence boundaries,* are indications of regional changes in discharge or sediment load brought about by tectonic or climatic forcing.

In a recent review, Holbrook and Bhattacharya (2012) carried out a detailed deconstruction of the concept of the subaerial erosion surface as a conventional sequence boundary; that is, as a surface that accords to the conventional concept of an unconformity as a time barrier. Their review presents additional detail and argument along the lines of that developed by the experimental work of Strong and Paola (2008), touched on above, and reprises and expands on the field evidence from incised valleys presented by Holbrook (2010), which is discussed below. In this writer's view, the subaerial erosion surface survives this examination

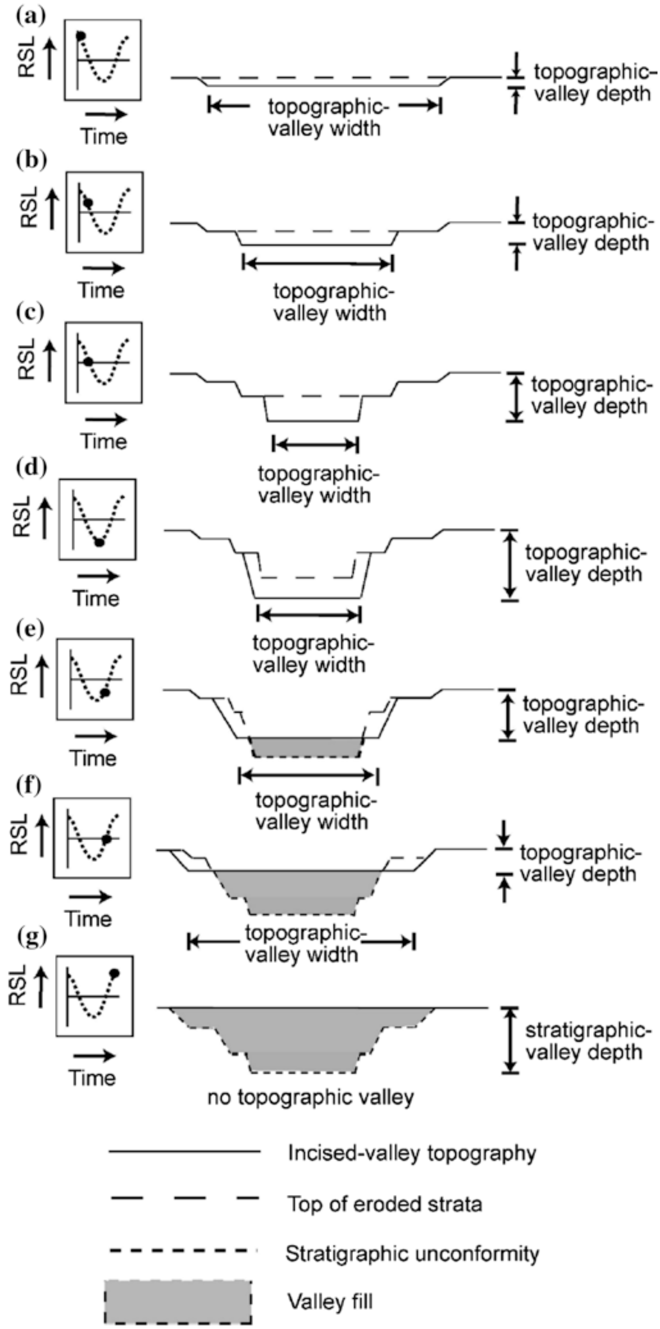


Fig. 6.14 Model of stratigraphic valley formation during a complete base-level cycle. Note the distinction between the topographic valley and the stratigraphic valley, the latter continuing to undergo modification until the end of the base-level cycle, by which time it is assumed to be completely filled (Strong and Paola 2008, Fig. 6, p. 589)

weathered, but matured, as a useful concept. However, Holbrook's (2010) work, in particular, presents a vista of time, particularly missing time, on multiple time scales, that is worth pondering at length by those of us who too readily toss around millions of years as unreducible elements of our geological histories.

6.3.3 Classifications of Erosional Sequence Boundaries

Incised valleys have received considerable attention from sedimentary geologists in recent years, in part because of their perceived significance in reconstructions of accommodation changes within a sequence context. It has also been recognized that incised valley-fills may constitute excellent stratigraphic traps for petroleum. Dalrymple et al. (1994, 2006) compiled two invaluable compilations of research papers on incised valleys, including a discussion of depositional models, and case studies of modern and ancient examples in a range of tectonic and climatic settings. Most of them contain estuarine sedimentary fills. Many other significant studies of fluvial valleys have been completed, as discussed below.

Incised valleys are difficult to study in the ancient record, because they typically exhibit a complex cut-and-fill stratigraphy that may be problematic to interpret with limited outcrop or subsurface data, and because of the huge range of scales that may need to be considered, from small, single-thread rivers to large braided systems or compound, laterally nested river systems (e.g., Holbrook 2001). Such complexities are well below the resolving power of most conventional chronostratigraphic techniques. In many basins, such as the low-accommodation settings that constitute the eastern flank of the Alberta basin, multiple cycles of base-level change have generated complexes of intersecting valleys which require special mapping techniques for accurate and reliable correlation. Examples from this and other areas are discussed below.

Plint and Wadsworth (2003, p. 1152) summarize the criteria used to recognize a valley-fill, as distinct from a simple, incised channel:

> (i) a valley is a linear erosional topographic structure, typically larger than a single channel form, that truncates underlying strata, including regional markers; (ii) the base and walls of the valley constitute a regionally mappable erosion surface across which there is an abrupt seaward shift in depositional facies; (iii) the erosion surface may be mantled by a pebble lag and/or characterized by a *Glossifungites* ichnofacies; (iv) the erosion surface should be traceable to an exposure surface, possibly characterized by a palaeosol on the adjacent interfluve; (v) depositional markers within the valley fill will onlap the valley walls; and (vi) tributary valleys are recognizable.

Attempts to develop wide-ranging descriptions and classifications of alluvial valleys were published by Holbrook (2001) and Gibling et al. (2011). Their key diagrams are shown here as Figs. 6.15, 6.16 and 6.17. Holbrook's (2001) classification (Fig. 6.15) is based on his studies of the mid-Cretaceous (upper Albian-lower Cenomanian) Muddy Sandstone of Huerfano Canyon in Colorado. He was particularly interested in the architecture of scour surfaces, channels and valleys, as expressed in outcrop over a wide range of scales, and was able to document

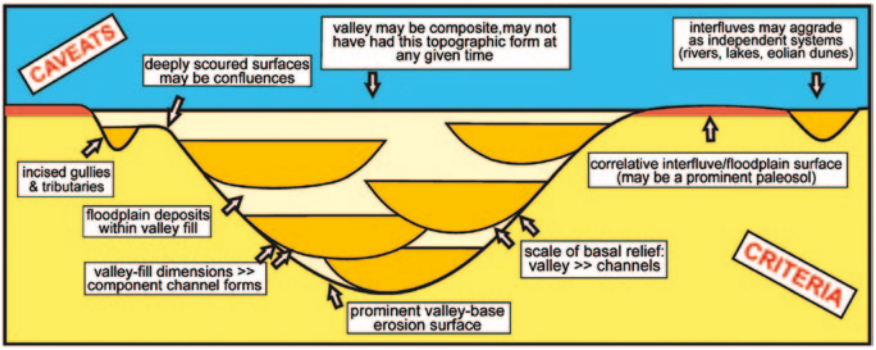


Fig. 6.15 The range of valley architectures, from simple to compound-complex, suggested by Holbrook (2001, Fig. 9, p. 211)

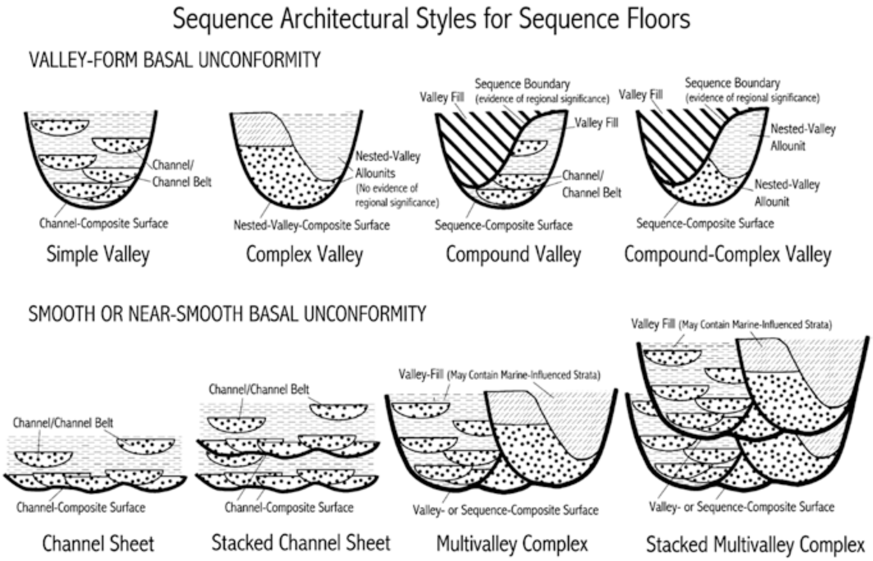


Fig. 6.16 Suggested guidelines for the identification of valley-fills in the ancient record. Caveats, in the *upper part* of the diagram, are features that may influence the decision to identify a valley-fill; criteria, shown in the *lower part* of the diagram, are features that may or may not be relevant in particular cases (Gibling et al. 2011, Fig. 9, p. 438)

these in some detail, based on the interpretation of photomosaic maps constructed from large and very well-exposed outcrops. He made use of the concept of hierarchical architectural elements and bounding surfaces developed by Miall (1985, 1988, 1996), defining surfaces in his project area of 6th, 7th, 8th and 9th-order rank, and suggested ways by which the larger-scale elements of this hierarchy could be classified using the principles of allostratigraphy, for example by assigning units such ranks as *allomembers* and *allosubmembers*. These ideas are integrated into the *Sedimentation Rate Scales* concepts in Table 6.1.

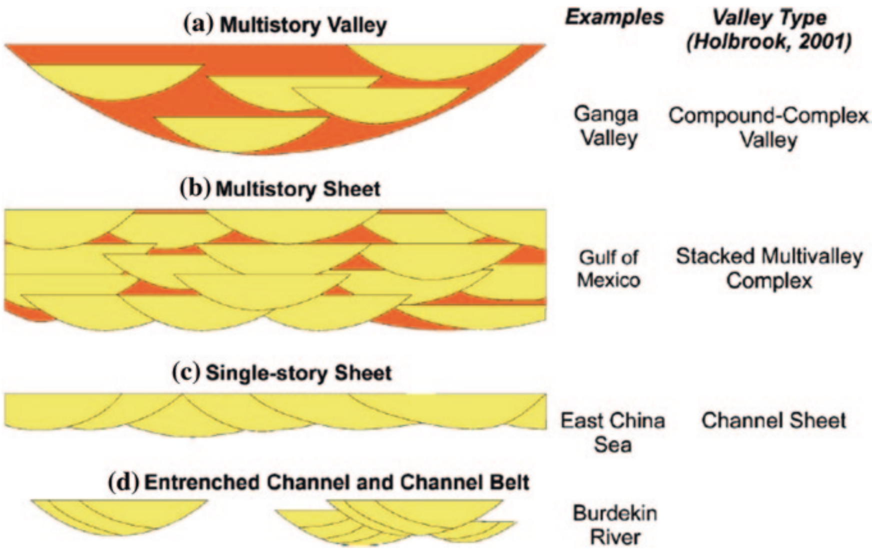


Fig. 6.17 A classification of Quaternary valley-fills, based in part on Holbrook (2001) (Gibling et al. 2011, Fig. 10, p. 439)

Holbrook's (2001) paper is an excellent illustration of how the architectural approach can be used to construct complex interpretations of a specific rock unit. However, caution needs to be expressed in generalizing from this study. As Holbrook says: "There is no compelling reason, however, to assume that 7th-order nested valley and/or 8th-order valley surfaces record preserved segments of previously continuous sequence bounding unconformities. Seventh- and 8th-order surfaces can just as well record valley-unique incision occurring where local climatic, tectonic and/or other drainage factors overwhelmed regional trends." (op. cit., p. 209). And again: "The possibility must also be considered that simple and complex valleys may potentially result from local incision and not be part of a basin-wide sequence boundary (op. cit., p. 210). Holbrook's (2001) study provides a good reminder of the need to keep time scales in mind when evaluating the ancient record against analogs from the late Cenozoic. A major sequence boundary typically constitutes a *SRS 9* or *11* event and may represent several million years of elapsed time. The processes of valley and terrace formation that have been observed in the post-glacial record (e.g., Blum 1993) typically represent the product of climatically-driven and autogenic events on a *SRS-8* or *9* scale, that is, on a time scale one to two orders of magnitude more rapid. Potentially, therefore, a major sequence boundary in the rock record could be the product of tens to scores of depositional and erosional cycles that have, at best, left highly fragmentary records. The intermediate scales represented by Holbrook's allowsubmembers are a particularly difficult scale to evaluate in the rock record, being typically too large to map reliably in outcrop, but too small to image on seismic records (except the type of high-resolution seismic used to map the shallow subsurface) or to correlate

Table 6.1 The hierarchy of channels and valleys at fluvial sequence boundaries

Type or rank (terminology from Holbrook 2001)	Bounding surface rank (Miall 1996)	Sedimentation rate scale (Miall, in press)	Time scale (years)	Processes	Driving mechanism
Nested channel cuts	4	5	10^0–10^1	Seasonal to 10-year flood	Normal climatic events
Channel scours	5	6	10^2–10^3	Channel avulsion	Autogenic valley-floor aggradation
Channel-belt (b allosub-member)	6	7	10^3–10^4	Channel-belt avulsion, river capture	Autogenic valley-fill; shifts in river due to climatic or tectonic events
Nested valley (nv allosub-member)	7	7–8	10^3–10^5	Channel-belt avulsion, river capture	Shifts in river due to climatic or tectonic events
Valley-fill	8	9	10^5–10^6	Channel-belt avulsion, river capture	Shifts in river due to climatic or tectonic events
Sequence boundary	9	9–12	10^5–10^7	Regional changes in ratio of accommodation to sedimentation	Major tectonic or climatic changes

using wireline logs. Terraces in river valleys are remnants of aggradational cycles that have been partly removed by subsequent degradation. As Archer et al. (2011) indicated, such deposits might constitute important, but overlooked, elements of a fluvial assemblage, and could provide useful clues to tectonic or climatic cycles, the evidence for which has been largely lost.

Holbrook's (2001) cautions regarding the possibly local nature of specific cut-and-fill architectures in the rock record has now been matched by the results of an important set of experiments that explored the development of incised valleys carried out by Strong and Paola (2008). They demonstrated the importance of a process that seems counter-intuitive: the shape of the final preserved stratigraphic valley is determined by erosion and sedimentation taking place during the rising limb of the base-level curve (Fig. 6.14). Pockets of channel deposits are formed by temporary deep scour throughout the base-level cycle, and some of these may be preserved as a result of lateral shifts and abandonment of the main fluvial channel. It should therefore be assumed that the basal fill of the valley is highly diachronous. It is entirely possible that such scour-fill pockets may be formed during the transgressive phase, towards the end of the phase of modification of the topographic surface, and

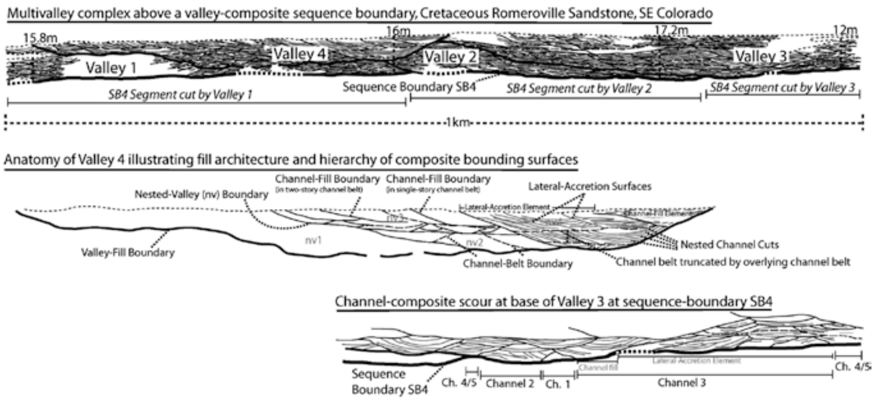


Fig. 6.18 *Top* An architectural interpretation of the Romeroville Sandstone (middle Cretaceous), Huerfano Canyon, Colorado, plus two close-up panels (*centre* and *bottom*) which illustrate enlarged portions of the profile. The numbered valleys in the top panel each constitute nested-valley complexes (Holbrook 2010, Fig. 1, p. 2; see Table 6.1 in this book for explanation of the terminology)

after the formation and burial of the base of the valley during the lowstand. In such cases, these deposits—the base of which defines part of the basal erosion surface and sequence boundary—would be younger than coastal deltaic or marine deposits formed at the lowstand; in other words, these beds, resting on the basal unconformity, are older than some of the beds cut by the basal scour near or at the coastline.

The diachronous, composite character of the basal sequence boundary is nicely illustrated in Fig. 6.18, which is from a Discussion of the Strong and Paola (2008) paper, submitted by Holbrook (2010). Holbrook illustrated the theme of diachroneity with a detailed architectural profile (Fig. 6.18), which shows a set of four valley-fills, each 200–300 m wide, that offset and nest into each other. Part of the basal sequence boundary is defined by the base of valley 1, part by valley 2, and part by valley 3. Each of the valleys, in turn, comprise fills formed largely by lateral accretion, indicating that the base of each valley is itself diachronous, on a smaller scale. This architectural profile illustrates the difficulty of distinguishing incised valleys from simple channels, because an incised valley is, in most cases, simply an amalgamation of channels that have formed slightly offset from each other over a period of time longer than that required for the formation of a simple scour by meander migration. Interpretations of the time scales associated with the channel and valley hierarchy illustrated in this diagram are provided in Table 6.1.

A further illustration of the ambiguity associated with sequence boundaries is illustrated by Figs. 6.19 and 6.20. The first of these is a detailed sequence correlation of a delta in the Ferron Sandstone of Utah. At first sight the incised valley-fill complexes (coloured red in Fig. 6.19) that mark the base of sequences 1 and 2 would appear to define a relatively simple, almost layer-cake stratigraphy. This is essentially the interpretation shown in Fig. 6.20a, in which all the incised valleys are assigned to sequence 2. However, there is no definitive proof that this is the correct interpretation. The regional erosion surface that marks the base of

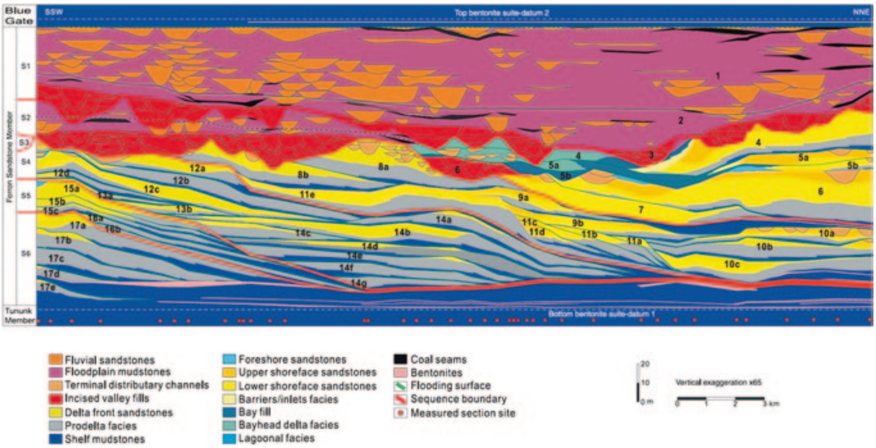


Fig. 6.19 Cross-section through the Ferron Notom delta (Turonian), Utah, hung on bentonites. Parasequences are numbered 1 to 17, and are grouped into six sequences. Two alternative Wheeler diagrams for this cross-section are shown in Fig. 6.20 (Bhattacharya 2011, Fig. 11, p. 135)

sequence 2 across the left-hand end of the section could be a composite of erosion surfaces cut at different times after the end of deposition of parasequence 7. This is the basis for the interpretation shown in Fig. 6.20b, in which it is suggested that the incised valley-fills that comprise the first valley-fill complex were formed at different times following the deposition of parasequence 9.

The most recent study of alluvial valleys, by Gibling et al. (2011), presented a set of criteria for recognizing valleys in the rock record (Fig. 6.16) and offered a classification based on Holbrook's work, with suggested Quaternary examples (Fig. 6.17). However, as noted in Sect. 6.3.1, the use of Quaternary examples as a basis for interpreting the ancient record may be questioned. We discuss below some of the specific examples cited in that paper, and reconsider them as interpretive tools from that perspective. In general, the applicability, and hence the value, of classifications (Figs. 6.15, 6.17), needs to be assessed against these important points:

- There exists a wide range of river channel sizes reflecting the wide range of rivers in nature. The hierarchy is not standardized according to scale (see Fig. 2.4).
- Valleys may vary in width not only according to river size but also according to the time they remain at the same stratigraphic level, leading to widening by lateral combing of the channels or by avulsive switching, as on many delta plains.
- Valleys of widely varying size may be developed within a small area, reflecting the presence of trunk rivers and tributaries of different sizes, the presence of deep scours at channel confluences, etc.
- Autogenic and allogenic processes, modified by lags and damping effects imposed by geomorphic threshold and response-time considerations, may develop comparable sedimentary responses that overlap and interact in unpredictable ways.

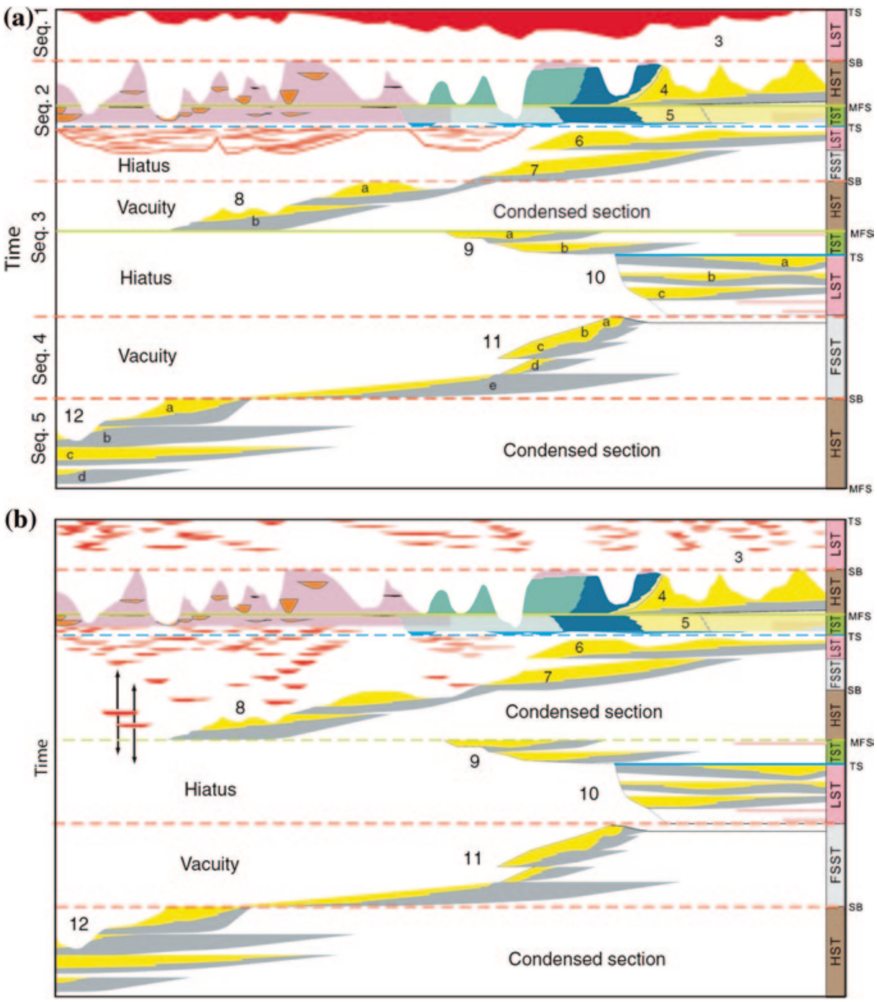


Fig. 6.20 Two alternative Wheeler diagrams for the cross-section shown in Fig. 6.19. The differences between them focus primarily on the chronostratigraphic interpretation of the incised valley-fills (Bhattacharya 2011, Fig. 14, p. 138)

- Major sequence boundaries, representing 10^5–10^6 years elapsed time, could encompass scores of higher-frequency aggradational-degradational events and cycles which have left extremely fragmentary to no record.

The conclusions from this discussion would appear to be that while classifications of valleys and valley-fills might constitute a useful descriptive terminology, considerable caution needs to be used in building generalized interpretations from these descriptions.

6.3.4 *Examples of Valley-Fills in the Recent and Ancient Rock Record*

Studies of incised-valley-fills in Alberta have a long history, because of their importance in the formation of numerous small- to medium-sized stratigraphic traps for petroleum. Three illustrations from a study by Wood and Hopkins (1992) are included here as Figs. 6.21, 6.22 and 6.23. In this area of low accommodation, large-scale fluvial systems appear to have repeatedly switched by lateral migration and avulsion, during recurring episodes of base-level rise and fall, building a complex suite of high-frequency sequences that cut into and intersect each other. Wood and Hopkins (1992) described techniques utilizing petrophysical data and detrital petrography to distinguish the valleys from each other, as an aid to wireline-log correlation. Chemostratigraphic techniques have been employed more recently for the same purpose. Chemical fingerprints for stratigraphic units may be derived by whole-rock geochemical analysis of core samples (Wright et al. 2010).

Arnott et al. (2002), Lukie et al. (2002), and Zaitlin et al. (2002) described the complex stratigraphy of the Basal Quartz Member of the Mannville Formation in project areas within the southern part of the Western Canada Sedimentary Basin. Accommodation in this area ranged from low to very low. At this time, southern Alberta was part of the foreland basin but was located some distance from the main locus of subsidence. The contemporary "deep basin" was centred in northeastern British Columbia (Zaitlin et al. 2002, Fig. 2). Subsidence trends and paleocurrent patterns indicate a strong influence of basement control, which Zaitlin et al. (2002) related to the heterogeneity the Precambrian crust under the influence of thrust-sheet loading. Tectonism triggered differential movement on structures in the basement generating responses that varied from region to region (Fig. 6.24). Stratigraphic studies indicate multiple phases of incised valley development

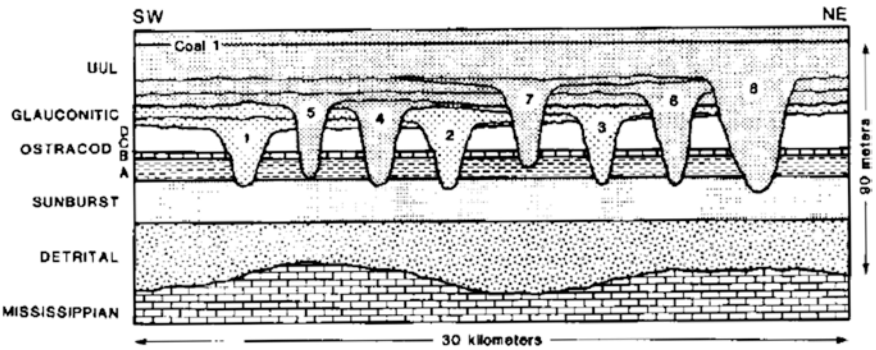


Fig. 6.21 Schematic stratigraphy of the Glauconite Member of the Upper Manville Group (Lower Cretaceous), Alberta, showing the succession of high-frequency sequences that intersect each other (Wood and Hopkins 1992, Fig. 5, p. 908). AAPG © 1992. Reprinted by permission of the AAPG whose permission is required for further use

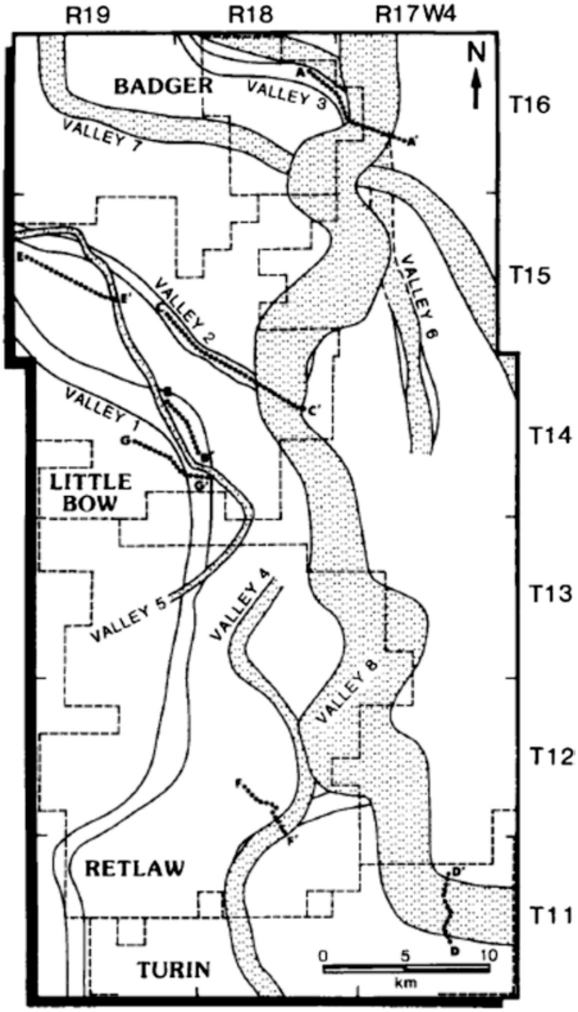


Fig. 6.22 Intersecting paleovalleys in the Glauconite Member of the Upper Manville Group (Lower Cretaceous), Alberta (Wood and Hopkins 1992, Fig. 4, p. 907). AAPG © 1992. Reprinted by permission of the AAPG whose permission is required for further use

(Fig. 6.25). Most would appear to be "complex valleys." In the Holbrook (2001) classification (Fig. 6.15) or "entrenched channel belts" in the Gibling et al. (2011) terminology (Fig. 6.17).

During the Early Cretaceous there was "an area of extremely low accommodation along the Saskatchewan-Alberta border and the SE corner of Alberta, where isopach values range between 0 and 40 m and net sedimentation rates [were] less than 2.2 m/my. This area was dominated by long periods of erosion and exposure, the development of paleosols, and multicycle incision of valley systems" (Zaitlin et al. 2002, p. 35). Accommodation increased slightly across a hinge-line defined by an aeromagnetic low, corresponding to a fault zone in the basement (Fig. 6.24). This maps as an east–west "hinge-line" in Fig. 6.26. To the north of

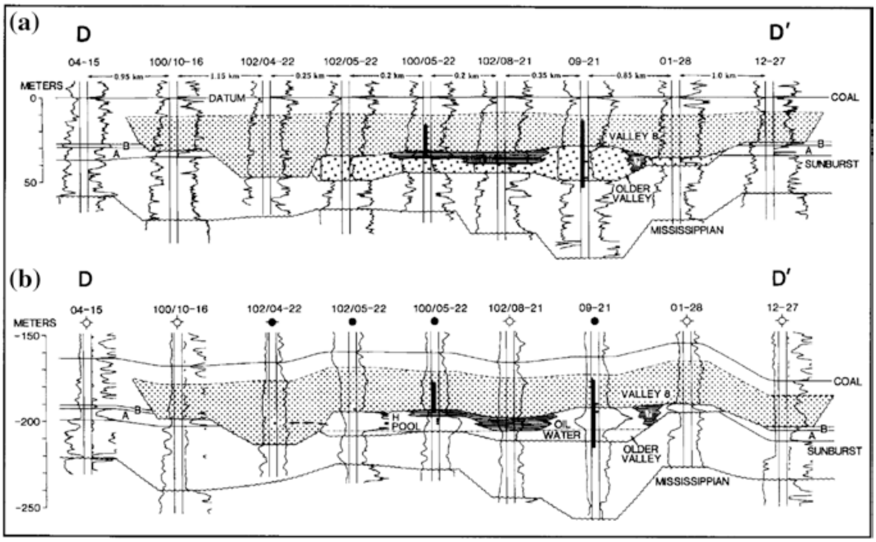


Fig. 6.23 Stratigraphic (**a**) and structural (**b**) cross-section through one of the valley-fill complexes illustrated in Figs. 6.21 and 6.22 other (Wood and Hopkins 1992, Fig. 17, p. 921). AAPG © 1992. Reprinted by permission of the AAPG whose permission is required for further use

the hinge-line is "an area of low-intermediate accommodation where isopach values range between 40 and 120 m, and net sedimentation rates ranged between 1.3 and 6.6 m/Ma. This area is characterized by mappable valley systems with sheet-like fluvial to coarse-grained meandering deposits, paleosols and thin coals at their base, changing upward into finer-grained meandering-fluvial to fluvial-estuarine systems" (Zaitlin et al. 2002, p. 35). The net sedimentation rate for the Lower Mannville was estimated as 6.6 m/my, and for the Upper Mannville 20 m/my (Zaitlin et al. 2002, p. 35). These are extremely low rates (10^{-3} m/ka), an order of magnitude less than *SRS-11*, and assigned to *SRS-12*.

An example of the valley system developed during one phase of the Mannville is illustrated in Fig. 6.26 and a stratigraphic cross-section through part of this system (section E-E′) is shown in Fig. 6.27. The "BAT" sandstones, like most of the valley-fill units, are prolific hydrocarbon reservoirs, forming numerous stratigraphic traps.

The Upper Cretaceous (Cenomanian) Dunvegan Delta complex of northwestern Alberta has been mapped in detail by Bhattacharya (1993), Plint (2002) and Plint and Wadsworth (2003). A cross-section through the complex (Fig. 6.28) shows a subdivision into seven regional allomembers, A to G, each representing a cycle of marine flooding, followed by progradation of river-dominated deltas. The deltas within each allomember consist of offlapping shingles, which represent individual deltaic lobes in plan view. Figure 6.29 is a map of the complete delta-plain and delta-lobe geography of Allomember E. The tributary systems and the trunk streams of this system have been mapped in great detail, based on the correlation of wireline logs in 4800 wells by Plint (2000) and Plint and Wadsworth (2003).

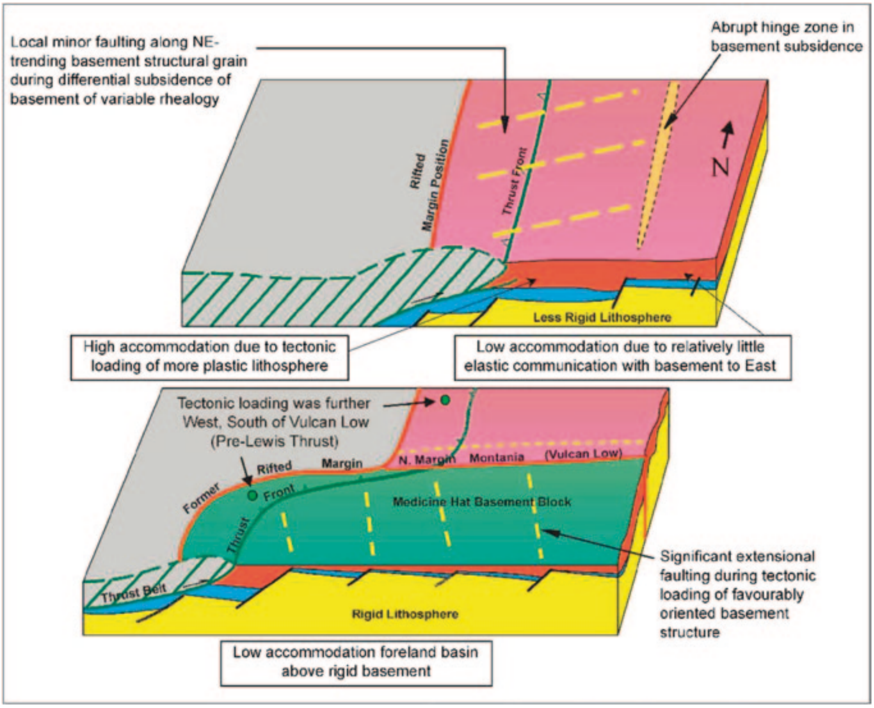


Fig. 6.24 Conceptual tectonic model for the Western Canada Sedimentary Basin in southwestern Alberta, showing the influence of Precambrian basement heterogeneity on the architecture and subsidence history of the foreland basin (Zaitlin et al. 2002, Fig. 34, p. 68)

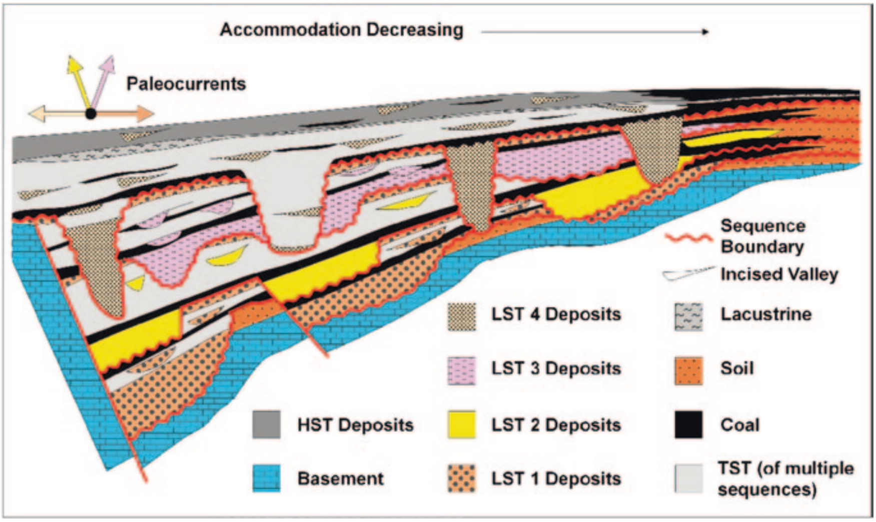


Fig. 6.25 Generalized stratigraphy of the Basal Quartz Member (Lower Manville Formation) as developed in a low-accommodation setting (Zaitlin et al. 2002, Fig. 3, p. 36)

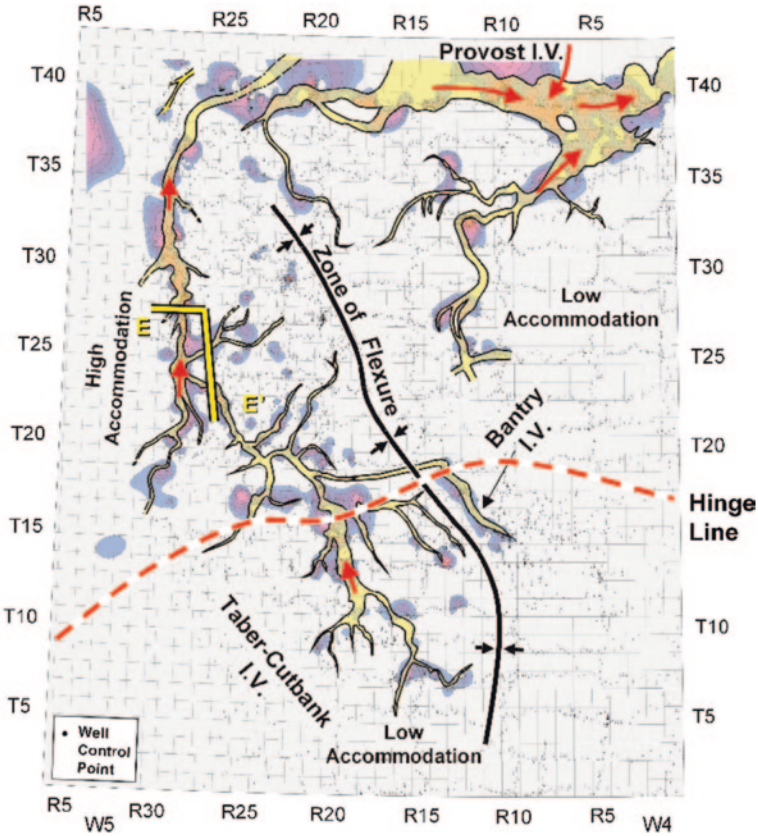


Fig. 6.26 Composite isopach and paleogeographic map of the BAT, a unit consisting of local sandstones named Bantry, Alderson and Taber. *Arrows* represent inferred paleodrainage direction. Where no contours are present, no BAT was deposited or preserved. Location of cross-section E-E′ (Fig. 6.27) is shown (Zaitlin et al. 2002, Fig. 21, p. 55)

The details for allomember E are shown in Fig. 6.30. Some of the valleys are up to 10 km wide, and can be traced for as much as 330 km in the subsurface. Like the Mannville valleys, these appear to be "complex valleys" in the Holbrook (2001) classification (Fig. 6.15) or "entrenched channel belts" in the Gibling et al. (2011) terminology (Fig. 6.17).

A typical strike-oriented cross-section through one of these paleovalleys is shown in Fig. 6.31. A schematic diagram illustrating the longitudinal stratigraphic relationships of a typical valley is shown in Fig. 6.32. The valley truncates deltaic shingles updip (to the north and west). In this illustration, three schematic gamma ray logs illustrate typical signatures through valley-fill deposits and host strata. "The centre log shows an upward-fining cap to the valley-fill, perhaps representing the muddy heterolithic part of a tidal point bar, or a tidal flat succession. The two flanking logs show at the top of the valley-fill, a sandier-upward succession that

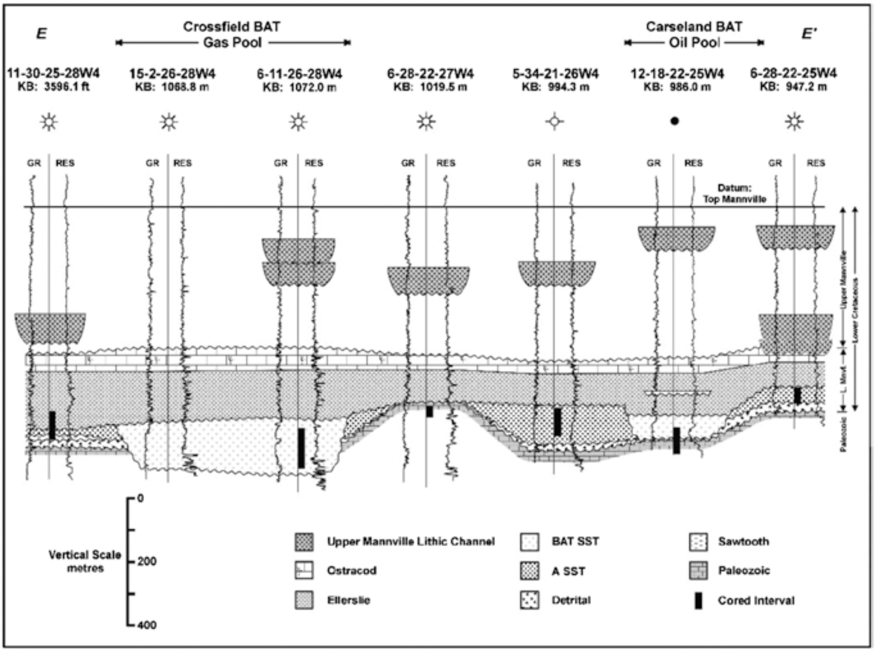


Fig. 6.27 Cross-section of the Mannville Formation, showing the low-accommodation BAT valley-fill geometry, and the more isolated channel architecture of the Upper Mannville along section E-E′ (location shown in Fig. 6.26) (Zaitlin et al. 2002, Fig. 22, p. 56)

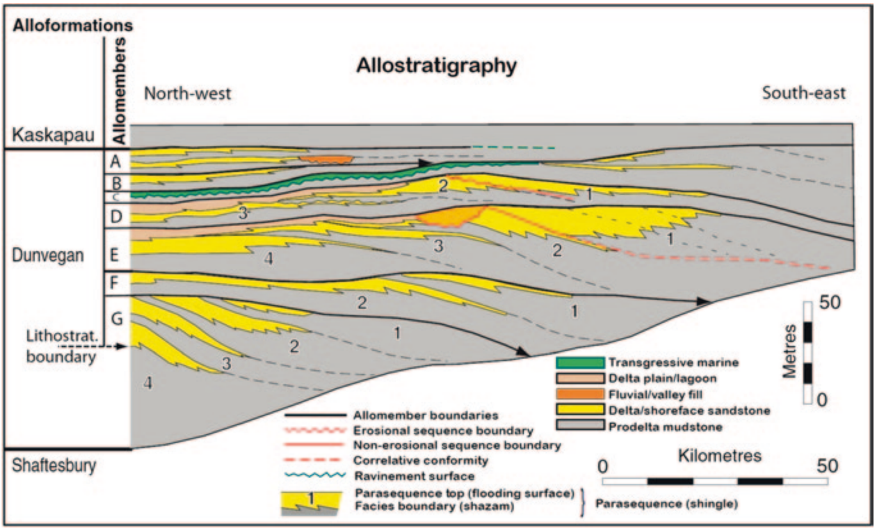


Fig. 6.28 Allostratigraphic subdivision of the Dunvegan Formation (Cenomanian) of northwestern Alberta (Bhattacharya 2011, Fig. 2b, p. 124; modified from Bhattacharya 1993)

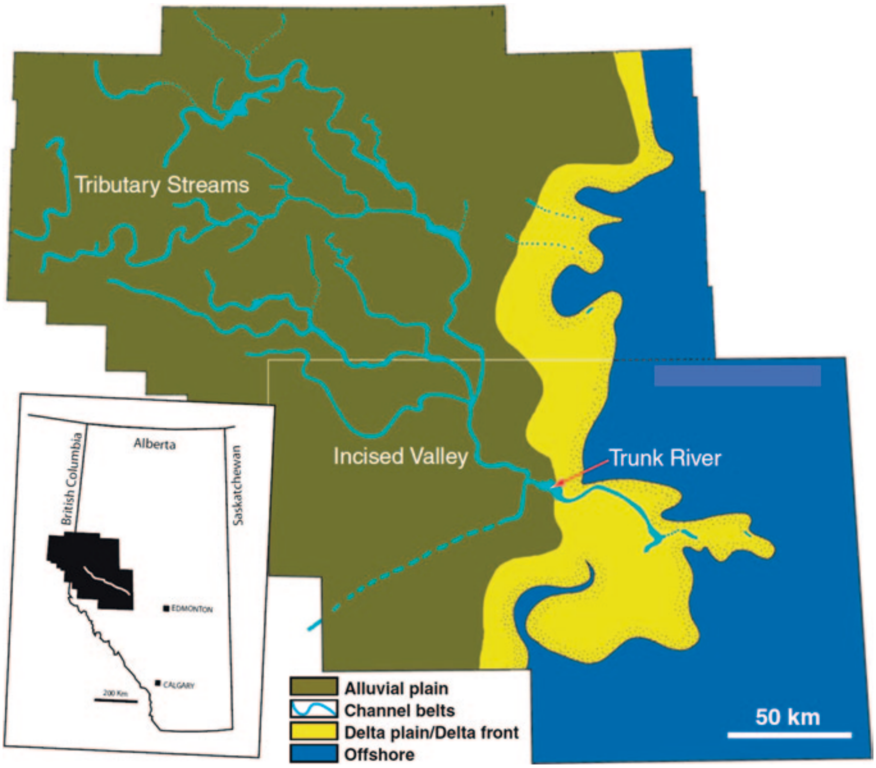


Fig. 6.29 The tributary and trunk river system feeding a river-dominated delta, Allomember E of the Dunvegan Formation, Alberta (see Fig. 6.28), based on subsurface mapping by Plint (2002) and Plint and Wadsworth (2003) using some 4,800 well records (Bhattacharya 2011, Fig. 24, p. 148)

might represent a bay-head delta. The muddy, estuarine portion of the fill appears to die out in the lowest reaches of the valley, which is interpreted as grading into a distributary" (Plint 2000, p. 293).

The Late Cenozoic incised-valley system mapped on the shelf of the East China Sea by Wellner and Bartek (2003) is cited by Gibling et al. (2011, Table 4) as an example of a "near-smooth basal unconformity channel sheet." It is a "multistorey sheet" in their classification (Fig. 6.17) and a "multivalley complex" in the Holbrook (2001) classification (Fig. 6.15). A map of the complex that developed during the MIS-2 isotopic stage (26–10 ka) during which sea level fell and rose about 100 m, is shown in Fig. 6.33, and a detail of the complex, imaged by the water-gun reflection seismic method, is shown in Fig. 6.34. In its entirety, the complex is more than 330 km across and reveals incision depths of up to 72 m. The authors attribute the width of the system to lateral channel migration. The detail of Fig. 6.34 reveals the presence of several separate channel systems, each of which was filled largely by lateral accretion. The shelf deposits into which this complex incised are marine deposits formed during the preceding highstand, and

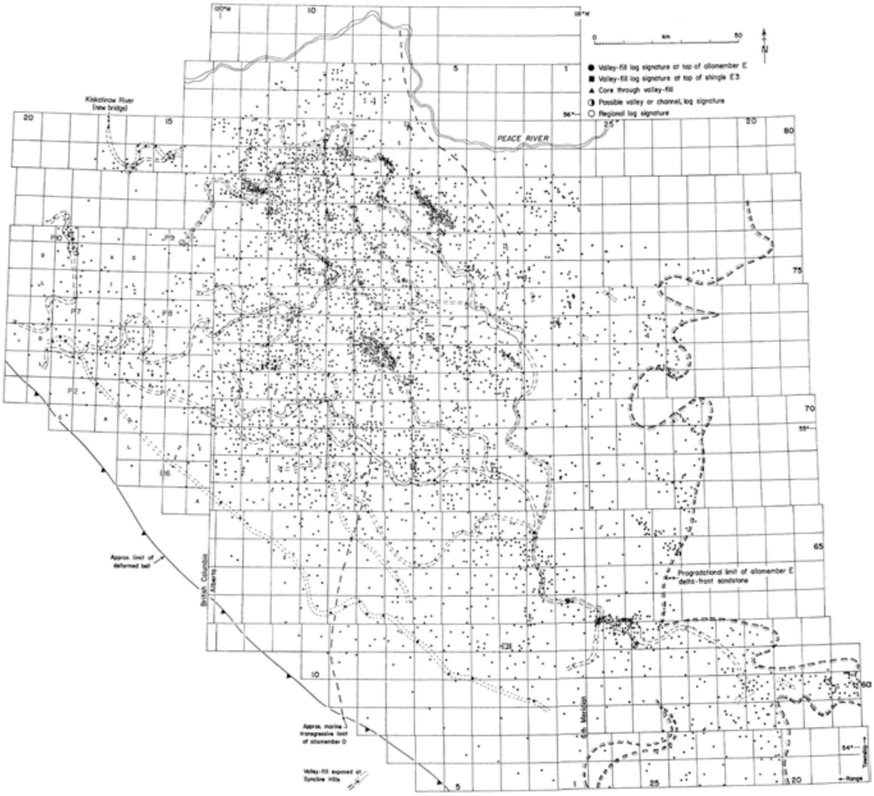


Fig. 6.30 The palleovalley system of Allomember E, Dunvegan Formation (the same system shown in schematic form in Fig. 6.29), showing well locations (Plint 2002, Fig. 5, p. 286)

they are overlain by the transgressive to highstand deposits of MIS-1. The appropriate sedimentation-rate focus for these deposits is *SRS-8*. Given that the entire complex is at less than 80 below present sea level, its long-term survivability through another full-glacial cycle is low. How relevant is this system, therefore, to the interpretation of the ancient record? It could be speculated that the more modest eustatic cycles predicted for parts of the Cretaceous, if applied to this paleogeographic environment, would have led to multiple re-incision of the same system, with many fragments spanning an age range of up to several million years that would become laterally amalgamated and with slow, steady subsidence, also offset vertically by a few metres. The end product might appear similar architecturally, but the total age range of the deposits might be entirely different.

Studies of the channel and valley complexes of the giant rivers of northern India have been undertaken by Tandon et al. (2006) and Sinha et al. (2007). Gibling et al. (2011, Table 4) suggested that they provide an example of a compound-complex system, in the Holbrook (2001) classification (Fig. 6.16). Tandon et al. (2006, p. 19) noted the difficulty of distinguishing "channels" from "valleys" in this area,

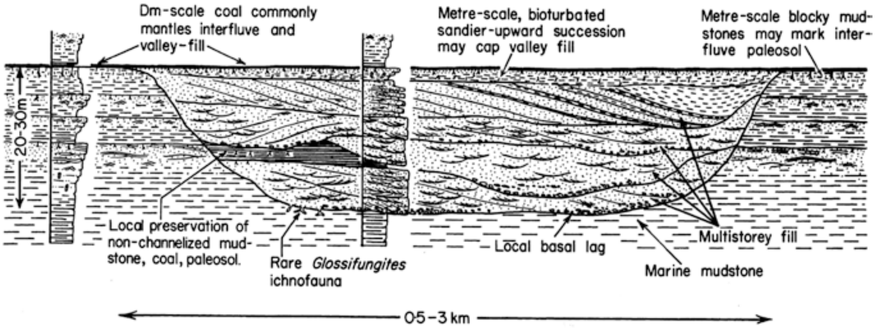


Fig. 6.31 Schematic longitudinal cross-section along a paleovalley in the Dunvegan Formation, Alberta (Plint 2002, Fig. 16, p. 293)

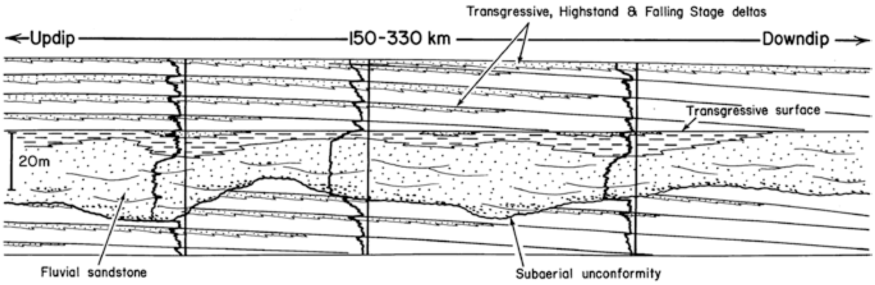


Fig. 6.32 Schematic strike-cross-section of a paleovalley in the Dunvegan Formation, summarizing the main sedimentological features of the valley-fill (Plint and Wadsworth 2003, Fig. 3, p. 1152)

given the large size of the autogenic channel scours, and the regular occurrence of deep scour caused by the seasonal monsoonal floods. Stratigraphic studies suggest that particular examples of deep incision can be related to monsoon events. Regionally, it is suggested that the major channels have shifted laterally as a result of the active tectonism in the basin, which is dominated by flexural subsidence and southward migration of the Himalayan thrust belt (Fig. 6.35). An example of this channel shifting is illustrated in Fig. 6.36, based on correlation of drill-core stratigraphy to the surface channel patterns. A large meander with a radius of about 10 km appears to have undergone a neck cut-off and the straightened channel also shifted southward, presumably in response to tectonism. It is not known what is the nature of the surface that defines the base of the compound valley interpreted from this stratigraphic evidence (the base of the grey area defining the valley in Fig. 6.36). It is likely that it constitutes a compound erosion surface, which may climb slightly through the floodplain-interfluve deposits in a southward direction.

Viewing the Ganga system through a speculative lens from the distant geological future, what is likely to happen to it over the next few million years? In other

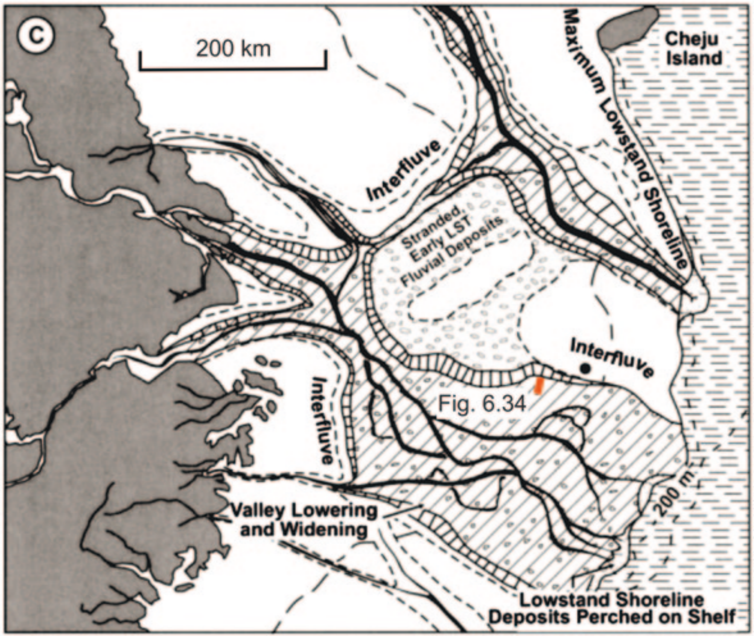


Fig. 6.33 Interpreted paleogeography of the vallely-fill complex of the East China Sea during the MIS-2 (late Wisconsin) stage. The location of the seismic section shown in Fig. 6.34 is indicated (Wellner and Bartek 2003, Fig. 9c, p. 937)

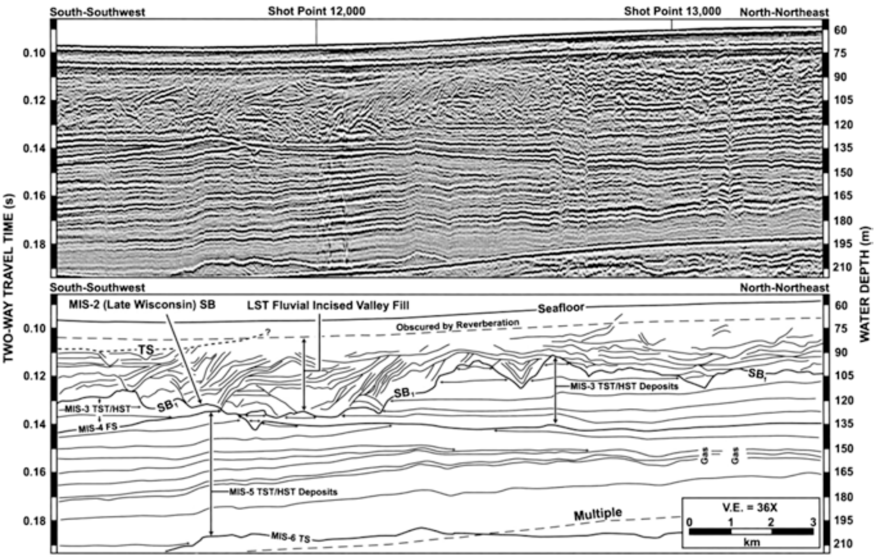


Fig. 6.34 Detail of the incised-valley system on the shelf beneath the East China Sea. Location is shown in Fig. 6.33 (Wellner and Bartek 2003, Fig. 5, p. 933)

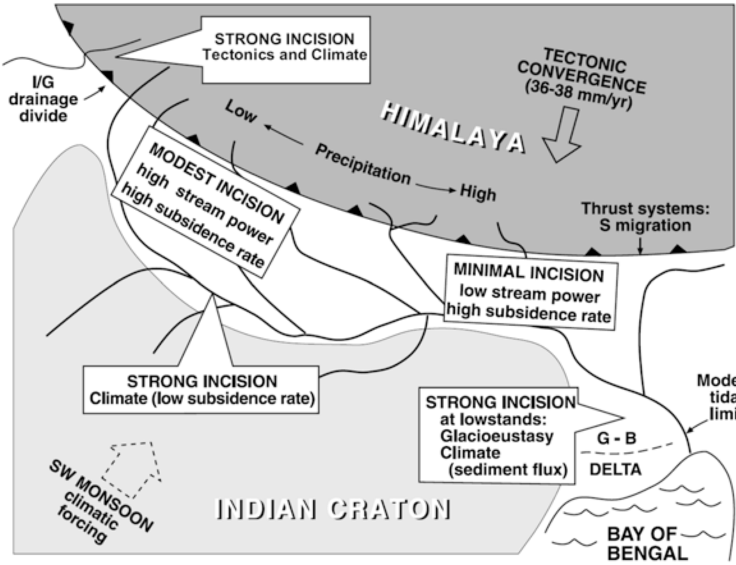


Fig. 6.35 Model for the dynamics of the river valleys in the Himalayan foreland basin (Tandon et al. 2006, Fig. 11, p. 29)

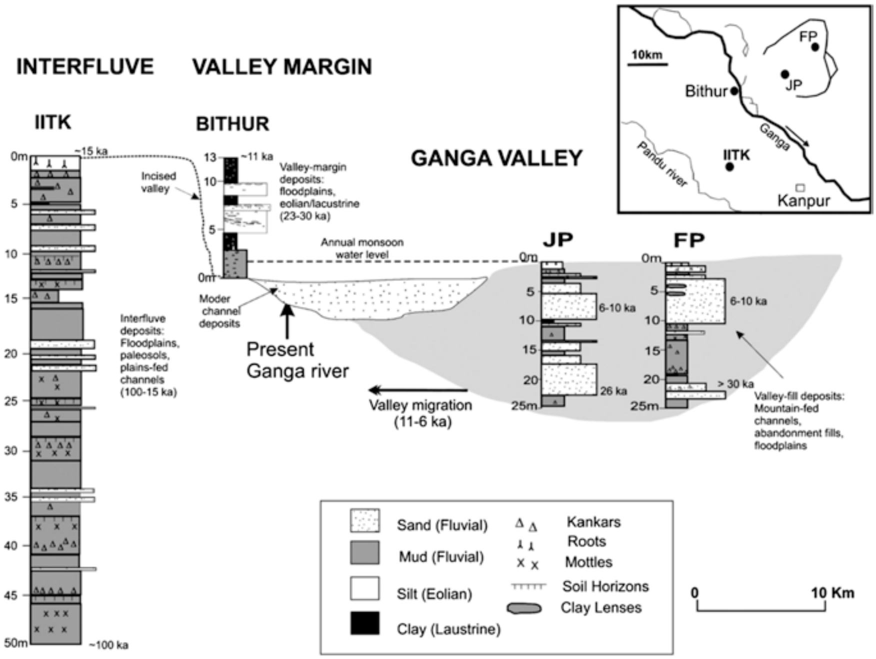


Fig. 6.36 Stratigraphy of the Ganga valley and interfluve, based on valley-margin outcrops and drill core. Dating of the channel fills in the drill-core sections suggests an episode of lateral migration of the Ganga valley about 30 km to the southwest (Sinha et al. 2007, Fig. 9, p. 405)

words, in what way do we need to adjust our interpretation of this system if we are to use it as an analog for interpreting the ancient record? High-frequency, high-amplitude glacioeustasy is not an issue in this case. Over an intermediate (10^4–10^5-year) time scale it is likely that monsoonal variability, autogenic meander shifting and avulsion would continually erode and replace specific channel fills and interfluve deposits, with only fragments becoming positioned for long-term preservation. It is likely that the southward shift of the system will continue. In a million years an active foreland may subside as much as 100 m (*SRS-10*: typical subsidence rate: 0.1 m/ka). Proximal deposits—those within a few tens of metres of the thrust front—may be tectonically disturbed, tilted, uplifted and partially eroded.

The examples of paleovalley described so far have been based on the availability of excellent subsurface control—tight networks of wireline logs, or high-quality reflection-seismic data, from which reliable dimensional information could be obtained. In the absence of such control, where the data consists of scattered outcrops or widely dispersed wells, it might be difficult to distinguish valley-fills from channels. Batson and Gibling (2002) described an example of where detailed facies and architectural methods could be used to distinguish valley-fills from simple channels and, employing additional criteria, such as associated facies and vertical profile, the different channel assemblages could be confidently assigned to different stages of base-level and climate cycles. The examples are from the Carboniferous, coal-bearing succession of the Sydney Basin, in Cape Breton Island, Atlantic Canada. Five distinct assemblages or "groups" of channel style were recognized, ranging from large incised systems, 4.6–13 m deep, to small channel sandstones less than 1 m thick. The large channels, assigned to group A, are interpreted as paleovalleys, filled during transgressive phases. This is suggested by the upward passage of the channel sandstones into finer-grained deposits, some showing evidence of marine influence, and into coals. The channels are commonly capped by beds containing hydromorphic soils.

6.3.5 Non-Erosional Sequence Boundaries

Sequence boundaries are a record of regional changes in depositional processes. The standard nonmarine to shallow-marine sequence boundary is based on the subaerial erosion surface, and commonly constitutes, or at least includes as part of its regional architecture, an incised valley system, as discussed at length in the previous section. However there are two mechanisms that may generate a nonmarine sequence boundary without evidence of significant erosion.

An incised valley is clear evidence for an episode of negative accommodation, which is usually equated to a fall in sea-level or to tectonic uplift. The key idea here is the change in depositional processes generated by allogenic forcing, and a reference to the buttress and buffer concept of Holbrook et al. (2006) may usefully be made at this point (Fig. 5.1). Significant changes in depositional processes may be forced by upstream controls, which cause a redefinition of the position of the buffer zone and the sedimentary processes taking place within it. Tectonic changes in depositional slope may change the calibre of the sediment load and modify the direction of tilt of the paleoslope; climate changes, through modifications to discharge or vegetation cover, may affect the balance between discharge and sediment load, leading to changes in the balance between aggradation and erosion (Fig. 5.16), and to changes in fluvial style. The result may be more-or-less synchronous, regional, mappable, changes in fluvial style with little or no evidence of significant erosion; in other words, a sequence boundary generated without the requirement for a period of negative accommodation. Sequences do not need to be defined on the basis of changes between high- and low-accommodation styles (whatever that may mean with respect to sedimentation rates, as discussed in Sect. 6.2).

The second mechanism for the generation of nonmarine sequence boundaries that lack evidence of large-scale incision is where forced regression takes place across a gently dipping continental shelf. If sediment supply is adequate, the river system will respond by aggradation of the coastal plain as it extends seaward in response to falling sea level. This mechanism was suggested by Miall (1991b), and the Canterbury Plains of South Island, New Zealand, provide an excellent example of the fluvial sequence stratigraphy that results (Browne and Naish 2003). The Canterbury Plains constitute a braid-plain some 50 km across (in the direction of depositional dip) and 200 km wide, formed from the coarse, gravel-dominated deposits eroded from the actively uplifting Alpine Mountains to the northwest (one of the gravel-bed rivers is illustrated in Fig. 2.13). During repeated cycles of sea-level fall, during the Neogene, the braid-plain extended some 100 km further seaward, to the shelf edge, and then underwent rapid transgression during the subsequent episodes of sea-level rise. This process has been repeated at least seven times (Fig. 6.37). The sheets of fluvial deposits so formed, constituting regressive systems tracts, consist of a predominantly gravel association, with minor sand and mud, comprising an assemblage similar to that described from the "Scott-type" of fluvial system by Miall (1996). Mutually-incised channel scours are present, marking the location of short-lived braid channels, but there is no evidence of deep incision. As the coast underwent regression during each episode of sea-level fall, the braid plain aggraded by developing very low-angle seaward-dipping clinoforms 10–30 m thick (Fig. 6.38). The clinoforms terminate in more steeply-dipping deltaic foresets on the upper continental slope, marking the lowstand. The subsequent transgressive deposits appear as hummocky or mounded features on high-resolution seismic records, and are interpreted as stranded beach, barrier and lagoon deposits formed during rapid transgression. The highstand deposits in the subsurface consist mainly of shelf muds. The modern coast is undergoing transgression along most of the Canterbury Plains shoreline, and is developing an erosional surface of wave ravinement (Leckie 1994).

The distinctive architecture of the Canterbury Plains deposits is a reflection of a very high sediment supply, steep fluvial slopes, and cycles of rapid accommodation changes during the falling stage, when the buffer zone underwent seaward translation, resulting in lateral and vertical expansion on the coastal plain. Once again, as in several of the other examples described in this book, the architecture is very

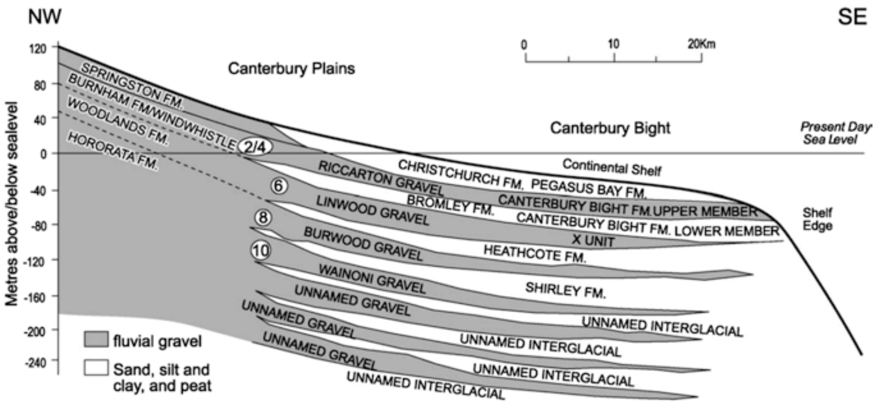


Fig. 6.37 Cross-section through the northern portion of the Canterbury Plains near Christchurch to the edge of the continental shelf showing the stratigraphy of alternating lowstand fluvial gravels and sands and highstand sand, silt, clay, and peat. Numbers refer to inferred oxygen isotope stages based on radiocarbon age dating (Browne and Naish 2003, Fig. 2, p. 670)

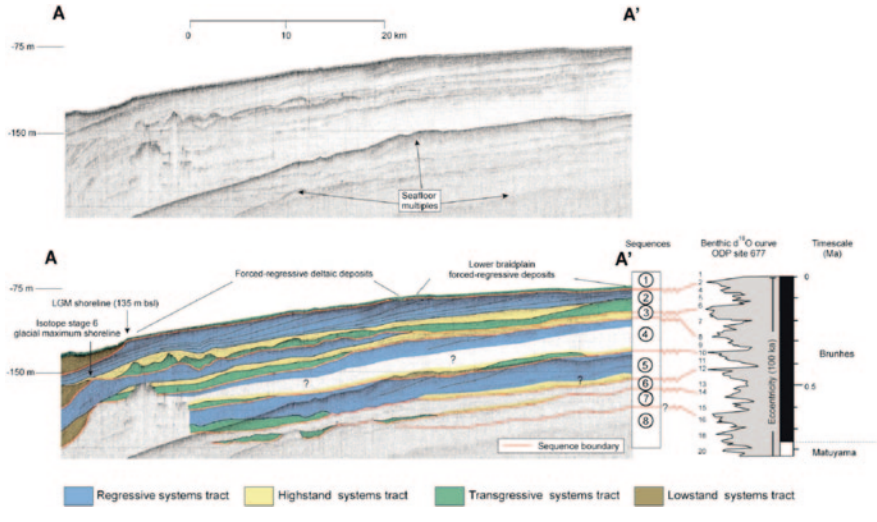


Fig. 6.38 High-resolution seismic section across the middle and outer portion of the shelf off the Canterbury Plains, correlated to the marine isotope stages based on limited ^{14}C dating, and age estimates based on sedimentation rates (Brown and Naish 2003, Fig. 17, pp. 69–70)

much a product of the late-Cenozoic high-frequency, high-amplitude changes in accommodation. Absent the high-frequency glacioeustatic cycles, how would the architecture of this coastal braid-plain be different? Assuming the same tectonically active source area and high sediment supply, it would seem likely that the main difference would be that, because of the much lower rate of accommodation generation, there would be a substantial rate of sediment bypass, with much of the coarse detritus delivered to the coast to be deposited as steeply-dipping deltaic clinoforms, and much of the sediment would then undergo slope failure and further transport deeper into the ocean by sediment-gravity flow processes. There is no reason to suppose that the base of the regressive braidplain systems tract would be bounded by deeply incised valleys or channels. It would not display a clinoform configuration, but would probably be seismically incoherent, reflecting an internal architecture of repeated channel avulsion and mutual erosion to shallow channel depths.

The Castlegate Sandstone (Upper Cretaceous) of the Book Cliffs in Utah contains a good example of what Miall and Arush (2001a,b) termed a *cryptic sequence boundary*. The base of the formation is defined by a major change in facies, but detailed mapping (Fig. 6.39) indicates that there is an unconformity which truncates the Buck Tongue (a shale unit) and defines an upper and a lower part of the unit, which therefore constitute two separate sequences (Willis 2000; Yoshida 2000). Tectonic control is clearly indicated by the regional shifts in paleoslope from one sequence to the next (Fig. 6.40). During periods of slow subsidence, basinward sediment transport was facilitated, following the antitectonic model of Heller et al. (1988), leading to the development of extensive, sheet-like depositional units. The lower Castlegate Sandstone and the Bluecastle Sandstone are examples. These deposits represent tectonically-generated nonmarine analogs of lowstand systems tracts. They extend more than 150 km down depositional dip from the type area. The sequence boundaries at their base may represent intervals of considerable erosion, as indicated by the regional mapping documented in Van Wagoner et al. (1990). The sequence boundary that truncates the marine shale unit constituting the Buck Tongue truncates tens of metres of

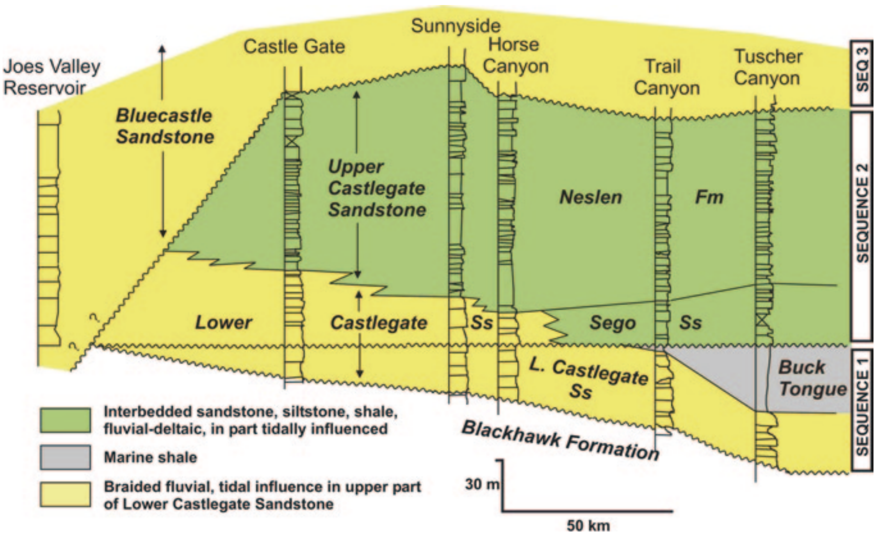


Fig. 6.39 Stratigraphy of the Castlegate Sandstone and related units, Book Cliffs, Utah. Note the presence of an unconformity capping the Buck Tongue and dividing the Castlegate Sandstone into two sequences (Miall and Arush 2001a)

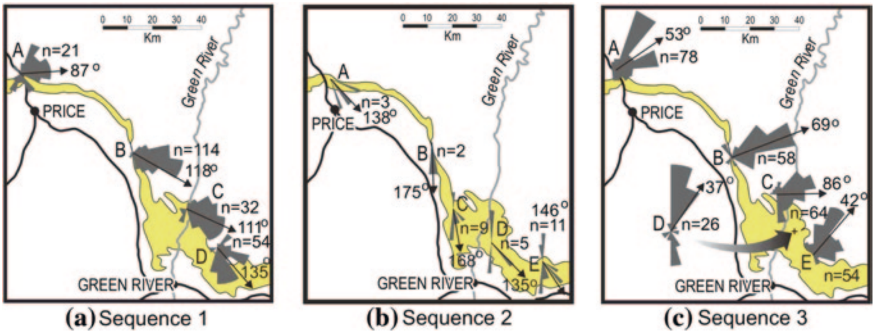


Fig. 6.40 Regional paleocurrent patterns in the three sequences of the Castlegate-Bluecastle succession, Book Cliffs, Utah (sequence stratigraphy shown in Fig. 6.39). Note the different regional tilts to the paleoslope indicated by crossbed orientations in each of the three sequences (Willis 2000, Fig. 17, p. 301)

strata in an updip (westerly) direction, and is difficult to detect at the type section of the formation (Fig. 6.41) because it superimposes similar facies across a contact that is no different on the outcrop scale from a typical channel scour surface. Miall and Arush (2001a) employed petrographic techniques to detect the sequence boundary, and determined that it probably was situated at surface D at the type section (Fig. 6.41), on the basis of changes in detrital composition of the sandstone across this surface, and evidence of early cementation of the beds below the surface, suggesting a considerable period of exposure and non-erosion of this surface during the accumulation of the Castlegate Sandstone.

In fact, as explicitly suggested by Bhattacharya (2011, Fig. 18), there may be multiple surfaces of regional erosion within the Castlegate Sandstone. Currently,

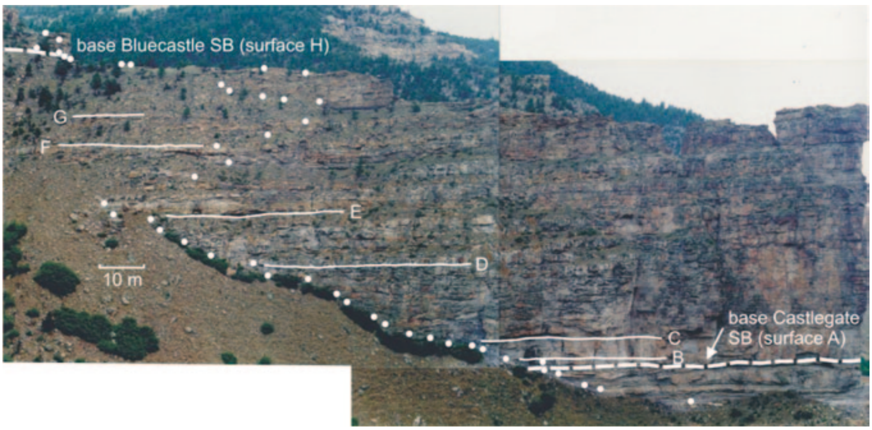


Fig. 6.41 The type section of the Castlegate Sandstone, at Price Canyon, Utah. Prominent channel scour surfaces, of 5th-order rank or higher in the bounding surface scheme of Miall (1996, 2010a) are indicated by heavy *white lines* and *letters*

mapping techniques are not adequate to the task of testing this more complex stratigraphic model.

6.3.6 Interfluves

Although sequences are conveniently defined on the basis of widespread erosion surfaces which, in the case of fluvial systems, commonly include incised valleys, there may be broad areas within fluvial systems, between the valleys, where little or no erosion takes place, and which therefore may not be identifiable on the basis of cut-and-fill architectures or major changes in clastic grain size (such as where a channel-floor deposit rests on a finer-grained lithofacies). These "*interfluve*" areas may be the site of significant soil development, and the identification and mapping of these may provide valuable additional data for sequence mapping and interpretation (Fig. 6.42). McCarthy and Plint (1998) provided the first important study of the use of paleosols in sequence identification and mapping, and have followed up with several additional detailed analyses (McCarthy 2002; McCarthy and Plint 2003), in which it is demonstrated how details of facies associations, paleosol type, and the microfacies and geochemical composition of the soils may be used to extend a sequence interpretation. For example, facies association and soil type may be able to help distinguish floodplain from upland, interfluve settings (Fig. 6.43).

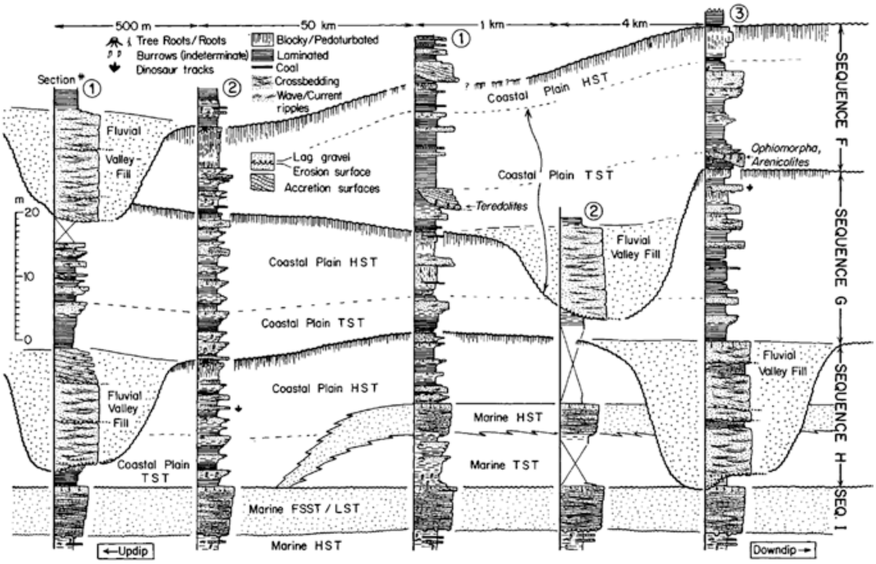


Fig. 6.42 The types of relationship between channel and floodplain clastic facies, and overlying pedofacies. Type 1–4 paleosols form in floodplain environments representing relatively high-accommodation settings, whereas type 5 paleosols form on interfluves during intervals of low to negative accommodation (McCarthy 2002, Fig. 5, p. 163)

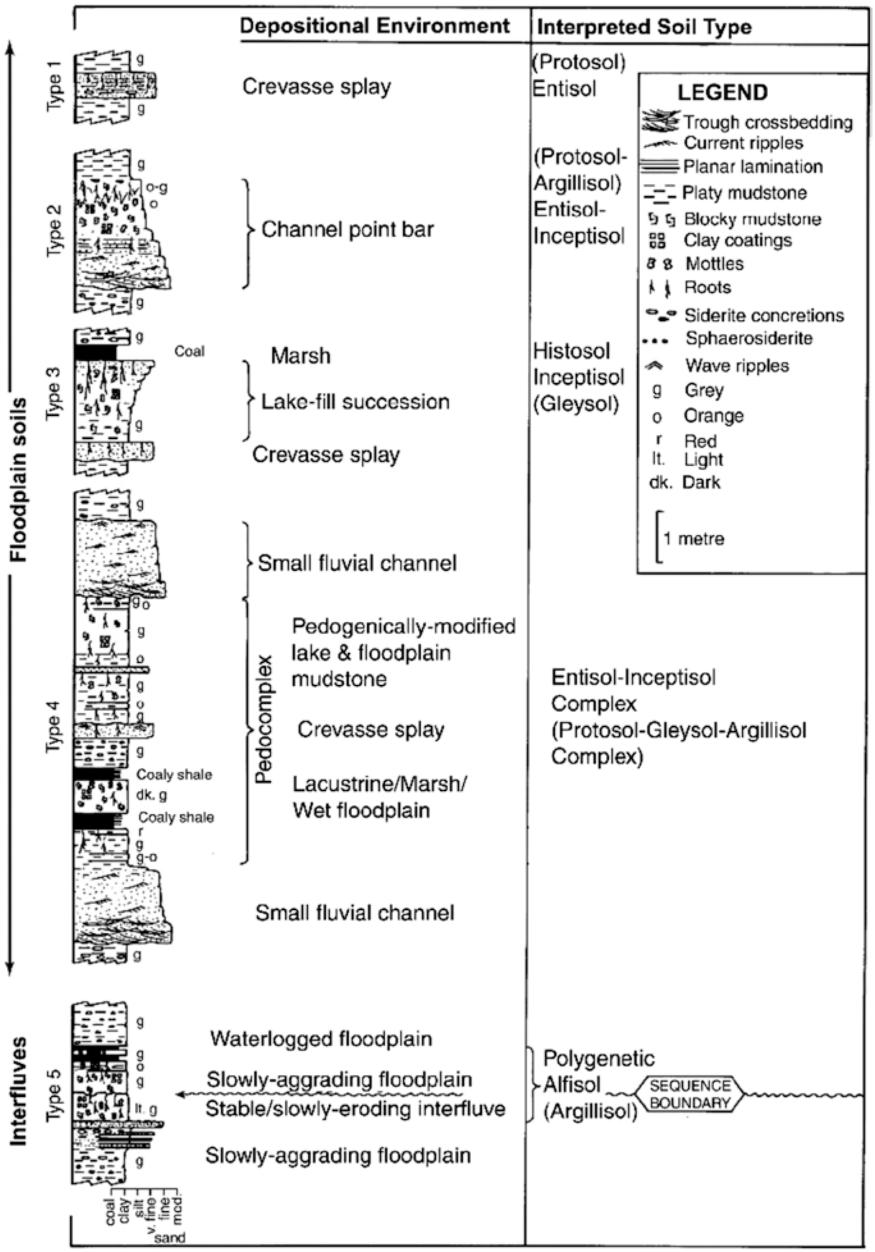


Fig. 6.43 An example of alluvial stratigraphy where sequence boundaries are marked both by incised valleys and by paleosoils (McCarthy and Plint 1998, Fig. 2, p. 388)

It should not be assumed that an interfluve marked by a paleosol represents a single cycle of slow to negative accommodation between cycles of fluvial aggradation. Well-developed soils require only a few tens of thousands of years to develop which, on the geological time scales that are normally involved in sequence analysis, means that there is ample time for several to many cycles of aggradation and degradation, and high-frequency climate change, each of which might have led to distinctive lithofacies and soil types deposited on the interfluve. As the studies referenced here demonstrate, some surfaces are marked by *cumulative paleosols* or *pedocomplexes*, formed by multiple cycles of soil formation, where several episodes of clastic or chemical sedimentation and minor erosion have also included periods suitable for soil development.

6.4 The Way Forward

Sequence concepts for fluvial systems have evolved significantly since the WMSM models were first proposed in the early 1990s, yet the speculative models of Allen (1974b) still have value as an intellectual exercise, constituting a set of thought experiments that explore the various controls that might affect alluvial systems. The basic base-level-change model, #5B (Fig. 6.1) remains a useful sketch for beginning to think about nonmarine sequence stratigraphy.

There is a considerable need for new field documentation of fluvial sequences. As this chapter has attempted to point out, the LAB and WMSM models, while valuable, may have reached their limit of usefulness in that they appear to be consistently directing researchers into developing interpretations at the wrong time scales. As noted earlier, these models are most appropriate for research focused at the *SRS 7* or *8* scales, at which scales autogenic processes are significant. The models are therefore of limited value in interpretations of the ancient rock record, most studies of which have been carried out at *SRS 9, 10* or *11*, where allogenic processes are predominant.

Future studies of the ancient also need to consider much more carefully the unique nature of the post-glacial record when using this record as a source of analogues for comparison and interpretation. The high rates of post-glacial sea-level change and the more complete preservation of events formed over the *SRS 1-7* time scales means that the Recent really is different, and to this extent uniformitarianism does not apply (Miall, in press). In several places in this chapter I have offered speculations about how particular modern depositional systems being used as analogs might have evolved under long-term conditions that lacked the high-frequency glacioeustasy of the late Cenozoic.

Holbrook's (2001, 2010) work on the architectural scale of fluvial systems, in particular his emphasis on the larger scale elements of fluvial systems, such as channel belts and valleys, has extended our thinking in an important way, by focusing attention on a scale of stratigraphy that is particularly difficult to come to grips with (see also Holbrook and Bhattacharya 2012). Lithofacies units and facies assemblages can be documented in outcrops and in drill core; sequences and systems tracts can be studied in reflection-seismic data and by careful correlation of wireline logs. However, this leaves a scale gap. Many units at the scale of the valley, typically hundreds of metres to a few kilometres across, and representing deposition over time scales in the range of *SRS 7* to *8*, are particularly difficult to capture systematically. The processes that take place at those times scales also tend not to lead to large-scale preservation, and the scales of data available to us do not make mapping simple. Geologists have long suspected that depositional fragments such as river terraces contain more information than can readily be extracted (e.g., Archer et al. 2011). Holbrook's examples from the Cretaceous of Colorado reveal how much can actually happen between the end of one cycle of base-level fall and the beginning of the next. Bhattacharya's (2011) alternative interpretations of the incised valley-fill units in the Ferron delta-plain (Fig. 6.20) make a comparable point. The simple sequence boundary at the base of sequence 2 in Fig. 6.19 may not be so simple after all. Sequence boundaries should perhaps be termed Rip Van Winkle Events. They may hide a lengthy series of events that have left little or no record, or one that is highly ambiguous in terms of temporal relations.

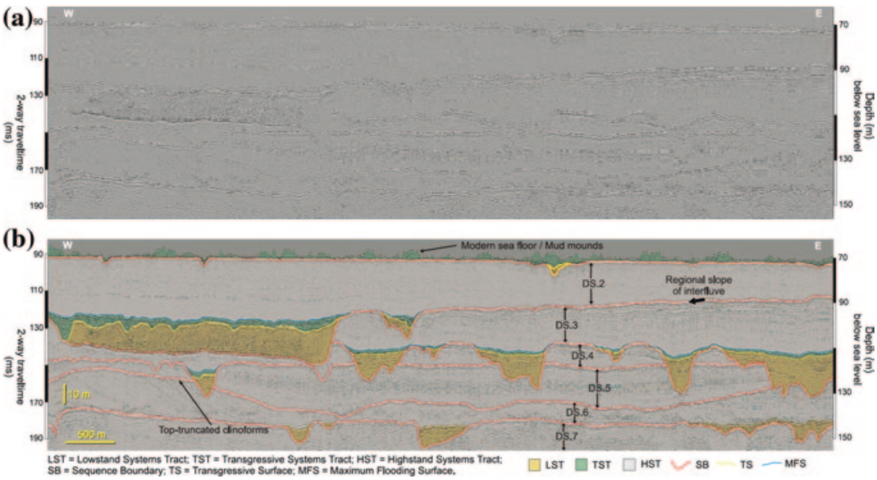


Fig. 6.44 The **a** uninterpreted and **b** interpreted high-resolution two-dimensional sparker cross section (peak frequency ~2000 Hz; tuning thickness ~25 cm [~10 in.]) showing six major stratigraphic discontinuities (marked in *red*) that define seven discontinuity-bounded stratigraphic (DS) units in the uppermost 80 m (262 ft) of sediments beneath the Gulf of Thailand Shelf. Location is given in Fig. 7.12, project #1. The valley fills, shown in *brown*, constitute the lowstand systems tracts of each sequence. The architecture of the valley-fills, as indicated by the seismic facies, reveals complex, overlapping lateral-accretion units. Another example of the valley fills, mapped using a series of time slices, is shown in Fig. 4.43. Beyond the valley margins sedimentation only occurred during the transgressive to highstand phase, when these systems were flooded by marine transgression. The highstand deposits consist largely of marine muds (from Reijenstein et al. 2011, Fig. 2, p. 1962). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

Modern data sets, particularly the advent of high-resolution, 3-D seismic, with its daughter product, seismic geomorphology, are providing many opportunities for more sophisticated interpretation. Some examples provided in Chap. 4 (Figs. 4.18 to 4.46) illustrate the types of information that may now be gleaned from the subsurface. Figure 6.44 is another example, showing the repetition of sequences 5–30 m thick generated by glacioeustatic cycles in the Gulf of Thailand. An incised valley and its tributary valleys from one of these sequences, imaged a few kilometres to the south, are shown in Figs. 4.41, 4.42 and 4.43.

Chapter 7
Large Rivers and Their Depositional Systems

7.1 The Definition of Large Rivers

The topic of big rivers has intrigued geologists for some time. The purpose of this chapter is to discuss the two major means by which the subject of big rivers and their associated depositional systems have been approached in the study of the ancient record, (1) Prediction from basin setting, including plate-tectonic environment, and (2) Analysis of the scale of depositional elements and other facies criteria, and study of sedimentary provenance.

There is a need, first of all, to agree on what is meant by the term "big river". Various authors have highlighted the length of the major trunk river, the area of the drainage basin, or the magnitude of the water or sediment discharge as primary criteria. The largest rivers on Earth are bigger than individual sedimentary basins (Ashworth and Lewin 2012, p. 86).

Potter (1978) was one of the first to examine large rivers systematically. His main objective was to explore the sources of the world's sandstone detritus and to understand the variations in its petrologic and geochemical composition. He noted that the drainage basins of the five largest present-day rivers (Amazon, Congo, Mississippi, Nile and Yenisei) amount to 10 % of the world's continental drainage area, the Amazon, alone, accounting for half of that total. Eleven of the largest rivers account for 35 % of the sediment load transported into the oceans. He noted that all but one of the fifty largest drainage areas of modern rivers had river lengths of more than 1,000 km, and that collectively these rivers drained 47 % of the total continental surface area, excluding the ice-covered regions of Greenland and Antarctica.

Schumm (1994), in a table compiled by B. R. Winkley, provided documentation of the basic size data for these fifty rivers; the shortest river listed is 900 km in length. A minimum of 1,000 km makes a convenient cut-off. Rivers of such dimensions average drainage-basin areas of 100,000 km^2. Gupta (2007), cited Hovius (1998) who documented the hydrologic geomorphic and climatic data for the world's 97 largest rivers, all of which had drainage areas greater than 25,000 km^2. Ashworth and Lewin (2012) noted several criteria that have been used to describe rivers as

A. Miall, *Fluvial Depositional Systems*, Springer Geology,
DOI: 10.1007/978-3-319-00666-6_7,

"big". Channel width has been used informally, suggesting that channels wider than 1 km "could reasonably be described as big." They noted some of the largest rivers on Earth have widths of up to 5 km (Amazon) or 10 km (the Brahmaputra with its complex of bars and islands). Ashworth and Lewin (2012) also noted that large rivers commonly widen and narrow and change style along their length in response to local geological conditions. Scour depth is also an indication of magnitude. The Brahmaputra (Jamuna) River locally scours to depths of 44 m (Best and Ashworth 1997); the middle Amazon to depths of 100 m (Ashworth and Lewin 2012, p. 85).

Amongst the examples of "big-river" deposits in the ancient record, a few stand out. The Hawkesbury Sandstone (Triassic) of the Sydney Basin, Australia, is a unit that has for some time been interpreted as the deposit of a large river (Conaghan and Jones 1975; Rust and Jones 1987), although other interpretations, including marine and eolian scenarios, have also been proposed, as summarized by Miall and Jones (2003). The main basis for the interpretation is the abundance of large-scale crossbed units, which are spectacularly exposed in numerous road cuts and cliff sections in the Sydney area. This example is discussed at greater length below, as illustrating the problems with interpretation from facies data.

The Athabasca Oil Sands of Alberta, Canada, are another deposit of a giant river system. Initially interpreted as the foresets of a delta (Carrigy 1971), Mossop and Flach (1983) demonstrated definitively that the distinctive, dipping strata exposed in the banks of the Athabasca River are the accretionary sets of giant, tidally-influenced point bars up to 25 m thick (Fig. 2.34). Much recent work has fleshed out this interpretation in detail (e.g., Fustic et al. 2008; Hubbard et al. 2011; See Figs. 4.30, 4.31).

A second Canadian example of a unit that has been interpreted on the basis of the scale of its component depositional elements is a sandstone of Neoproterozoic age in northwestern Canada (Rainbird 1992). Compound crossbed sets are up to 8.5 m thick and have been traced for up to 5 km. They suggested channel depths of 8.5 m within a braidplain system 150 km wide. Provenance data (discussed below) suggested a source on the other side of the continent.

Archer and Greb (1995) speculated about possible "Amazon-scale" drainage paralleling the Appalachian orogen during the early Pennsylvanian. Their primary evidence lay in the mapping of incised paleovalleys 5–10 km wide and up to 62 m deep, cut into Mississippian strata. The fill of these paleovalleys consists of quartzose sandstone and pebble conglomerate interpreted as the deposits of large braided rives. Based primarily on paleogeographic arguments the authors suggested longitudinal river systems extending from their mouths at the margin of the Ouchita Basin northeastward as far as Maritime Canada, or even Greenland. However, no architectural data, such as large bar forms or individual channel cutbanks were described that would provide evidence of the actual channel scale. As noted below, distinguishing the cutbanks of individual large channels from the similarly deep valleys generated by the incision of rivers that are not necessarily "big" is a perennial problem in fluvial sedimentology.

Many other examples of the deposits of large rivers in the ancient record are described and illustrated by Fielding (2007) and Fielding et al. (2012).

7.2 Tectonic Setting of Large Rivers

The focusing of water and sediment along a few exceptionally large drainages is an outcome of tectonic activity. But exactly what tectonic processes lead to big rivers is not at all a simple matter. Obviously, large areas of exposed continent are required, but beyond that, no simple, generalizations are possible. Dickinson (1988) attempted to classify continental drainage systems into four broad types, only the first of which seems guaranteed to generate long rivers with potentially large sediment loads. This is the "*Amerotype*" of continent, which is characterized by marginal orogenic belts, leading to asymmetric drainage patterns, with large rivers draining across the interior to a distant continental margin. The Amazon and Mississippi-Missouri systems fit this categorization. Dickinson's second category, "*Eurotype*" continents, are those which contain interior orogenic belts, from which rivers flow transversely along structural grain or out across marginal cratons. Asia displays both type of river system, the Tigris-Euphrates, Indus, Ganges–Brahmaputra, Irrawaddy, Mekong and Red rivers all flow along structural grain, while the great rivers of Siberia (Ob, Yenisei, Lena) are, in tectonic setting, comparable to the Mississippi and Amazon, in flowing from orogenic highlands out across vast, interior, cratonic platforms. Less predictable are Dickinson's third type of continent, the "*Afrotype*", characterized by centrifugal flow from rift highlands, or along rift axes. The Niger-Benue, Orange, Congo and Zambesi follow the first of these patterns, whereas the Nile fits, in part, the second. Dickinson's fourth type of continent, "*Austrotype*", are low-lying landmasses with no dominating orogenic highlands, and limited water and sediment. Australia, with only one significant river, the Murray, fits this categorization.

Many other workers have reviewed the origins and tectonic setting of large rivers (Miall 1981, 2006b; Hovius 1998; Tandon and Sinha 2007; Fielding 2007). Miall (2006b) suggested that there are five tectonic settings in which big rivers are particularly likely to develop (Fig. 7.1): (1) Axial flow along foreland basins, including retroarc and peripheral basins (e.g., Indus, Ganges–Brahmaputra, Tigris-Euphrates); (2) Strike-slip basins (e.g., Red River of Vietnam); and (3) Rift basins (e.g., Nile, Rhine, Rio Grande); (4) forearc basins (e.g., Irrrawaddy); (5) Accretion of large terranes in complex orogens can also create conditions for large rivers to develop, commonly following tortuous paths around and across sutures (Danube in Europe, Columbia and Fraser in North America). Fielding (2007) suggested that big rivers develop primarily in three settings: (1) marginal to contractional orogenic belts, where they are oriented along or close to the locus of maximum subsidence parallel to the structural grain; (2) Within major rift zones, where they flow along the axis of the zone, from rift basin to rift basin. Some of the basins may be occupied by lakes, through which the river flows; (3) Extending outward from the centre of large, stable cratons.

Miall (1981) developed a tectonic classification of rivers and drainage patterns, pointing out the two main common patterns of flow, longitudinal (or axial) and transverse, relative to structural grain. Longitudinal flow is most likely to lead to

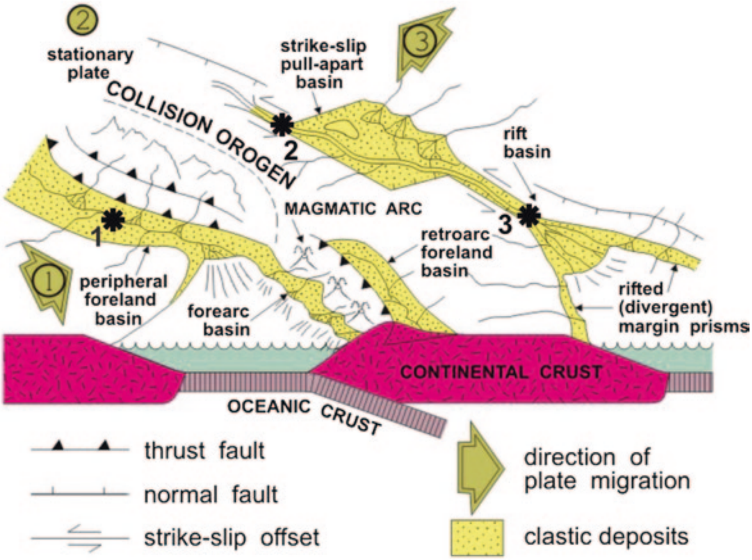


Fig. 7.1 The tectonic setting of alluvial basins. The *asterisks* locate the tectonic settings where some of the largest rives typically occur: (1) longitudinal trunk rivers in foreland basins (e.g., Ganges, Indus, Tigris), (2) rivers aligned along the axis of pull-apart basins, (3) rivers flowing along the length of major rift systems (e.g., upper Nile)

big rivers, except where orogens are flanked by wide cratons, across which the flow of large rivers could be classified as transverse with respect to the first-order tectonic fabric (although both of the most typical large, cratonic rivers cited above, the Mississippi and Amazon, are guided in their lower course by the ancient structural fabric of the basement). None of these generalizations is reliable, even for analyzing and classifying present-day rivers, and should be taken as providing only the most generalized of guidelines for interpreting the ancient record.

Many big rivers appear to follow no obvious or simple rules. For example, the Don-Volga system of Kazakhstan flows *towards* the Alpine-Himalayan orogen, draining into the Caspian Sea. Some rivers seem to make no sense at all, at least in terms of elementary patterns of tectonic evolution (e.g., Colorado of the Colorado Plateau in the United States). In such cases, a knowledge of the principles of antecedent drainage is required to understand the present day course of the river. In North America, late Cenozoic continental glaciation, which paid no regard to the tectonic framework, had a profound effect on river systems. The Mississippi is as big as it is primarily because of the southward diversion of drainage from what is now the north-central plains of the United States (Knox 2007). The Mackenzie, of northern Canada, might appear to be a typical longitudinal river flowing along the axis of the foreland basin flanking the Cordilleran orogen, but is, again, a product of ice damming and diversion of rivers that formerly, and probably throughout much of the Cenozoic, flowed eastward toward Hudson Bay (McMillan 1973;

Duk-Rodkin and Hughes 1994), but became diverted northward along overflow valleys into the ice-free corridor between the Cordilleran and cratonic ice caps. Brookfield (1993, 1998) showed how large rivers change in size and shift their course in response to regional tectonism.

7.3 Prediction and Analysis of the Deposits of Big Rivers

7.3.1 Facies Criteria

Since the relationship between fluvial hydraulics and bedform generation was first elucidated in the late 1950s (Middleton 1965), it has become an elementary component of sedimentological investigation that the scale of depositional elements in channel systems is in some way related to the scale of the depositing system. Early work by Allen (1965a, b, 1966) included the development of two themes, (1) ideas about the relationship between the size of bedforms, such as dunes, and the depth of the flow in which they formed, and (2) recognition of the significance of lateral accretion in channels opened a means to estimate the scale of channels from the thickness and dip angle of the accreting sand body.

Ashley (1990) compiled much useful information about bedform scale and its relationship to channel size, and there is now a significant body of literature on this topic, much if it summarized in recent books by Miall (1996) and Bridge (2003). It is now generally understood, following Ashley's (1990) compilation, that mesoforms (dunes) are typically scaled to the size of the channel in which they form. Where crossbed set thickness is consistently greater than about 1.5 or 2 m, it has long been assumed that this indicates the former presence of a deep channel. For example, this was one of the most obvious criteria that first led sedimentologists to suggest a "big-river" origin for the Hawkesbury Sandstone of the Sydney Basin, Australia (see Rust and Jones 1987 and Miall and Jones 2003, for a review of the sedimentological analysis of this famous and spectacular deposit, which is discussed further in Sect. 7.4). Leclair and Bridge (2001) suggested that water depths could be estimated from the thickness of crossbed sets, based on a compilation of the hydraulic relationship between these parameters as observed in modern rivers. However, the preservation of bedforms as crossbedding in the rock record may entail significant erosion of the top of the bedform by the turbulent scour cells that precede the migrating foresets, and so this relationship should only be regarded as a qualitative guide. As applied to the analysis of core or microscanner data, this method also suffers from the problem that it may be extremely difficult to accurately determine crossbed scales from the limited intersections of the 1st-, 2nd-, or 3rd-order bounding surfaces that are encountered in single vertical transects through a deposit. Ashworth and Lewin (2012) pointed out that in larger rivers, the response of dunes to major changes in stage may not be predictable, so that the relationship between dune height and flow depth may have considerable scatter. Leclair (2011) demonstrated that flood stage in large rivers may leave no clear sedimentological signature.

The relationship between channel-fill dimensions and channel scale has long been pursued in the sedimentological literature, following Allen's (1965a, b) preliminary ideas. Statistical relationships for the point bars of meandering rivers were developed by Leeder (1973), and much geomorphic work on channel dimensions was introduced to the geological literature by S. A. Schumm and F. G. Ethridge, notably in their 1978 review article (Ethridge and Schumm 1978). By far the most reliable indicator of large channels is the presence of very large inclined strata, which form by lateral accretion across the full depth of a channel. The thickness of the point bar is therefore a measure of the bankfull depth of the channel. These earliest attempts to reconstruct fluvial paleohydrology focused on the scale of "point-bar" deposits—now referred to more generally as lateral-accretion deposits. More recently, the development of the two- and three-dimensional approach to the study of outcrop facies architecture (architectural-element analysis of Allen (1983), Miall (1985, 1988)) has added new tools for the determination of channel scale, while the parallel work of Bridge (2003), combining studies of modern rivers with numerical and theoretical modeling, has added another set of methodologies for the analysis of the ancient record. Gibling's (2006) compilation of data from the rock record is now the definitive study of channel and sandbody scale.

The use of two- and three-dimensional facies data is important for this work, because the vertical scale of a depositional unit is not a sufficiently definitive item of evidence. Channel-fill scale is a good indicator of channel depth, but this relationship can only be employed if the geologist is certain that the observations relate to the autogenic fill of individual channels. (Leeder 1973; Ethridge and Schumm 1978). As illustrated in Fig. 2.17, there may be cycles within cycles in any given fluvial deposit, and determining which if these (if any) provide accurate indications of channel depth may be no easy task. The presence of lateral and downstream accretion deposits can be used to identify individual channels. Examples are discussed below. Fielding (2007, Fig. 7.3) illustrated several examples. Blum et al. (2013) demonstrated that point bar dimensions (thickness and width) scale with the size of the river based on such indicators as drainage area and discharge (Fig. 7.2).

In the ancient record, multistory channel fills are common, and comparable successions, showing the typical upward fining grain size distribution of an aggrading channel, may be developed by allogenic processes; for example, the tectonic cyclothems of Blair and Bilodeau (1988) (see Fig. 5.2). It is necessary, therefore, to distinguish between the laterally-limited deposits of channel fills and the regional lithostratigraphic units that reflect regional allogenic processes. It is also important to take note of such issues as the response time of fluvial systems to allogenic forcing, and the style of autogenic cyclicity that may be triggered by allogenic episodes (see Kim and Paola 2007, and Sect. 5.2.1). Another possible source of confusion is the potential similarity between the deeply scoured channels of large rivers and the deeply entrenched incised valleys filled with the deposits of smaller-scale river systems that form during episodes of negative accommodation. Fielding (2007) and Gibling et al. (2011) discussed the characteristics of incised valleys and the criteria that need to be assessed in order to

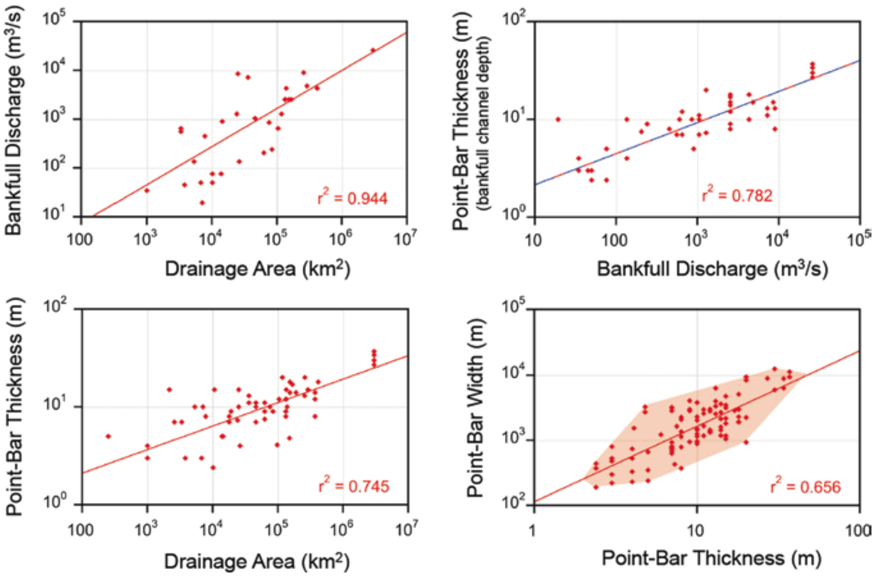


Fig. 7.2 Relationships between drainage-basin area, discharge and point bar scale for present-day large rivers (Blum et al. 2013, Fig. 5, p. 135)

correctly identify them. Fielding (2007, pp. 105–106) suggested the following diagnostic criteria for recognizing incised valleys:

> (1) the basal erosion surface that records the lowstand of relative sea level (the "sequence boundary") must be regional (basin-wide) extent; (2) facies associations overlying the basal erosion surface should be markedly different from those below the surface; (3) the erosional unconformity should remove underlying strata, which may be preserved beneath interfluve areas of the sequence boundary; and (4) incised valley fills have a distinctive internal architecture that is commonly multi-storey, and that records the progressive rise in base-level through the filling of the valley. The first of these criteria is crucial and probably the only one that is truly diagnostic. This criterion is also very difficult to satisfy unequivocally, however, because it requires that any correlation of a surface between outcrops or subsurface data points demonstrates absolute equivalence.

7.3.2 Trunk Versus Tributary Systems

River systems typically consist of a central trunk river with many tributaries supplying sediment and water from the edges of the catchment. Most river systems are predominantly "tributive" in nature, that is, in tracing a river upstream from its mouth, tributaries join the trunk river and the trunk river may eventually diverge into several smaller rivers of comparable size. Each of these, and each of the tributaries, also show patterns of branching and divergence in an upstream direction. Important exceptions to this general pattern commonly occur at basin margins where there may be a major contrast between a high-relief watershed area and the more subdued topography of the alluvial plain. In this setting, steep, sediment-laden streams emerging from the upland source area may diverge to form fan-shaped distributary patterns at the break in slope between the steep, confined configuration of the mountainous valley and the much flatter, less confined or even unconfined configuration of the plain. Such distributary systems are common elements of preserved ancient fluvial systems, occurring as, for example, coarse alluvial fan deposits banked against a basin-margin fault or prograding from an active thrust front. At their mouths, rivers are also commonly distributive in nature, where the trunk river splays out as the distributaries of a delta.

This elementary dissertation on fluvial geography is a necessary introduction to a debate that has been taking place in recent years regarding the relative importance of tributative versus distributive (or contributary versus distributary) patterns of fluvial channel patterns, as preserved in the rock record. In a pair of provocative essays, Weissmann et al. (2010, 2011) claimed that "*distributive fluvial systems* (DFS) commonly called megafans, fluvial fans, and alluvial fans in the literature, dominate the fluvial depositional patterns in active continental sedimentary basins, with a much smaller proportion of the fluvial area in these basins covered by rivers that are generally confined." (Weismann et al. 2011, p. 327). They suggested that trunk rivers may be confined at the centre of a basin between DFS actively prograding from opposite basin margins, and that this could account for their supposedly volumetrically minor importance in the rock record. They also claimed that such DFS are poorly represented by descriptions in the sedimentological literature.

These authors further claimed that most modern concepts relating to facies characteristics, facies models and architectural studies as employed in the interpretation of the ancient record were largely derived from studies of modern tributaries in degradational settings, and this led them to ask the question "are these the appropriate rivers to study when attempting evaluate facies distributions in the rock record?" (Weismann et al. 2011, p. 327).

To take their last point first, their question addresses one of the fundamental assumptions underlying the sedimentological method, the method of analogy that bases interpretations on the uniformitarianist principle that "the present is the key to the past." Virtually all of the fine details of the facies analysis method are based in whole or in part on observations and interpretations made from comparable environments and processes that can be observed in modern settings. Elsewhere Miall (in press) has addressed this question in general, as it relates to the stratigraphic and sedimentological interpretation of all clastic deposits in shallow-marine and nonmarine environments. It is suggested that processes that are completed and can be observed within a human time scale (*Sedimentation Rate Scales 1–6*, see Table 2.1) may be legitimately compared to preserved sedimentary products in the ancient record that developed over those time scales. Examples would be bedforms and many of the macroforms in fluvial systems. Whether a modern river is in a degrading setting or not is irrelevant to the development of bedforms and macroforms, because the degradational time scale (>10^4 years) is so much greater than the time scale of macroform and bedform generation. In the case of longer-term products, such as channel complexes and deposits on a sequence scale (*SRS 7-12*), issues of preservation need to be considered, and here the problem of interpretation of fluvial systems is part of a much broader issue that has not been fully addressed in modern sedimentological and stratigraphic studies. Miall (in press) described the work of the "geological preservation machine" which, by definition, has not completed its work in the case of modern deposits, because the main geological processes that involve long time scales (sea-level change, subsidence, climate change) are, at all times, in the middle of cycles that require at least 10^4 years for completion. In Sect. 6.2 I discuss the application of sequence models for fluvial systems, which have been developed primarily from studies of the modern record, to the interpretation of sequences in the ancient record.

As to the first question, that of the relative importance of distributive fluvial systems (DFS), the papers by Weismann et al. triggered a vigorous response from Fielding et al. (2012) who disputed the claim by Weismann and his colleagues that DFS predominate in the rock record. This is an important debate, because it bears on the mappability and predictability of fluvial systems in the subsurface. At least some of the examples cited by the latter do not appear to support their thesis, based on their own data. For example, the upper Brahmaputra River of northeast India (Weismann et al. 2011, Fig. 4E) is clearly a tributary system.

The main criteria for DFS are (1) radiating distributive channels, (2) downstream decrease in channel size, (3) downstream decrease in sediment grain size, and (4) channels are largely unconfined. Fielding et al. (2012) suggested that Weismann et al. (2010, 2011) consistently applied only the first of these criteria in the identification of their examples. Many are illustrations of dryland settings, where the DFS is essentially a terminal fan grading basinward into lower energy environments.

The most important counter argument to the importance of DFS is the abundant documentation of the deposits of large rivers in the rock record, as discussed above. Numerous examples of large rivers, including some illustrated by outcrop photographs of large-scale lateral-accretion deposits, were supplied by Fielding (2007) and Fielding (2012). The two worked examples described later in this chapter also illustrate the significance of the "big river" in two different stratigraphic settings, where there is no evidence for a DFS. In future, the increasing use of high-quality seismic-reflection data and the employment of the methods of seismic geomorphology will make resolution of this question simple.

The facies distribution and sedimentological composition of the deposits of large rives depends to a considerable extent on the pattern of tributaries and the sediment supplied by them (Ashworth and Lewin 2012). Figure 7.3 illustrates some of the variations that are present on Earth at the present. Some rivers are supplied primarily from a distal, mountainous source (e.g., Amazon, Indus: Fig. 7.3a,c); longitudinal rivers, such as the Ganges, may receive most of the sediment from tributaries flowing in primarily from one side (Fig. 7.3b). Where the discharge and sediment load of the trunk and the tributaries is very different, the fluvial style may vary markedly across a basin (see Fig. 2.11 and discussion thereof). Some large rivers, such as the upper Nile and the Danube, flow across tectonic barriers separating successive

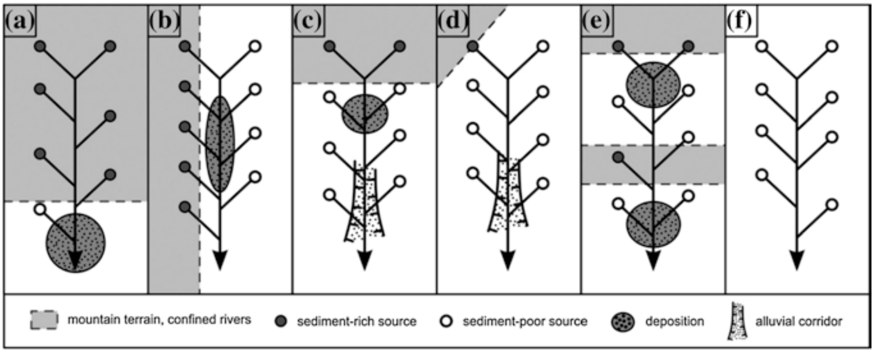


Fig. 7.3 Types of trunk-tributary configuration, in relation to areas of sediment sourcing and sediment deposition. **a** = Mountain-dominated (e.g., Amur, Mekong), **b** = Lateral tributary-dominated (e.g., Ganges, Mississippi, Paraná), **c** = Headwater-dominated with foreland depositional basins (e.g., Amazon, Orinoco), **d** = Headwater-dominated with lowland alluvial corridor (e.g., Ob, Mackenzie), **e** = Alternation between depositional basins and mountain belts (e.g., Danube, Yangtze), **f** = Few mountainous sources (e.g., Congo, Rio Xingu). From Ashworth and Lewin (2012, Fig. 2, p. 87)

basins along their course. Tandon and Sinha (2007) provided a similar classification. They distinguished between mountain-fed, foothills-fed, plains-fed and mixed-fed river systems. Mountain fed rivers are likely to carry significant coarse bedload, whereas plains-fed rivers may be dominated by suspended load, with the result that the deposits that become part of the preserved stratigraphic record may show very different facies and architectural characteristics in different parts of the basin. One of their illustrations (Fig. 7.4) shows how there may a wide range of facies and architectures in such settings. Large channel bodies form where a major fan extends across the basin margin; smaller rivers yield smaller channel units, interbedded with abandoned-channel deposits, some including peat/coal deposits. Interfluve areas may lack coarse (conglomerate, sandstone) bodies altogether.

Ashworth and Lewin (2012) defined four main types of sedimentary-basin landscapes that are characteristic of large fluvial depositional systems (Fig. 7.5). The first type (a) is displayed by rivers with abundant wetlands, such as the Magdalena, the second (b) by larger rivers sourced from distant catchments with few major tributaries, as along much of the length of the Mississippi. Rivers with many tributaries (c), such as the Ganges and Indus display main channels with numerous tributaries that may be of similar or quite different fluvial style. The fourth style, the bedrock channel (d) occurs in the geological record as an incised valley (Fig. 7.5).

7.3.3 Provenance Studies

Provenance studies can demonstrate distant sediment sources, but they cannot indicate the size of the river. Nevertheless, this approach has generated useful supplementary data for paleogeographic analysis and, in several cases, this has had

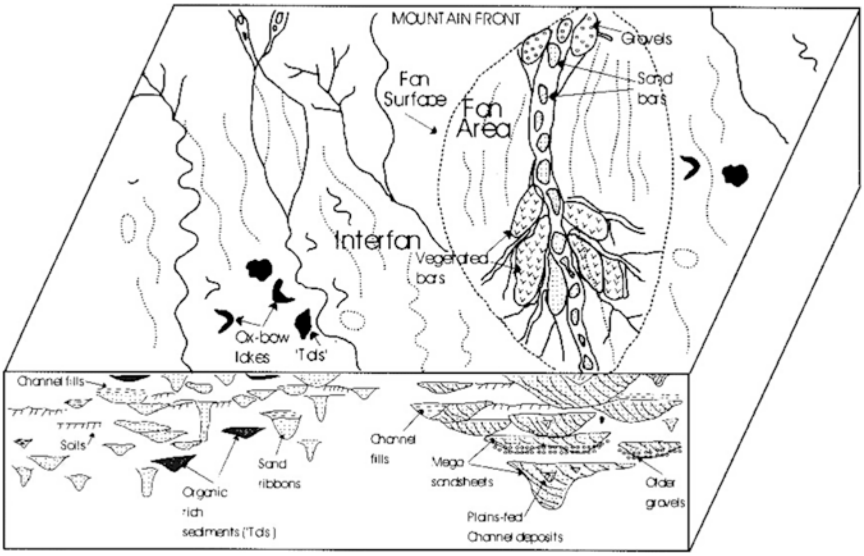


Fig. 7.4 Architectural cartoon of a typical suite of transverse rivers entering a basin from an orogenic highland, showing the variety of architectural styles to be expected in this type of geological setting (Tandon and Sinha 2007, Fig. 2.7C, p. 19)

significant implications for the reconstruction of continental-scale tectonic evolution. Three examples from North America are briefly mentioned here.

Rainbird et al. (1997) used U–Pb and Sm–Nd geochronology to explore potential sediment sources for the Neoproterozoic sandstone earlier interpreted as the deposit of a large braided river system. The ages of the majority of the zircons separated from this sandstone indicate Grenville sources. Rainbird et al. (1997) argued that paleogeographic characteristics of the sandstones, including the regional structural setting and paleocurrent data, suggested sources to the southeast, and they concluded that the Grenville orogen of eastern north America, some 3,000 km distant, was the most likely source for the bulk of the sand.

Having speculated about continental scale provenance controls (Dickinson 1988), Dickinson subsequently explored in detail the provenance of the exceptionally thick and extensive Permian and Jurassic eolian sandstones of the southwestern United States (Dickinson and Gehrels 2003) using a similar approach to that of Rainbird et al. (1997). Zircon ages indicate a variety of Precambrian sources, and the authors concluded that the Appalachian orogen was the major sediment source for this sand, indicating long-distance transport across the Laurentian craton. Neither the location nor scale of specific river systems could be identified by this research, but given the distance of transport and the volume of sediment displaying these provenance characteristics that is now located on the western continental margin, it seems highly probable that big rivers were involved. The westerly tilt of the craton during the late Paleozoic and early Mesozoic could at least in part be attributed to continental-scale thermal doming leading up to the rifting of Pangea,

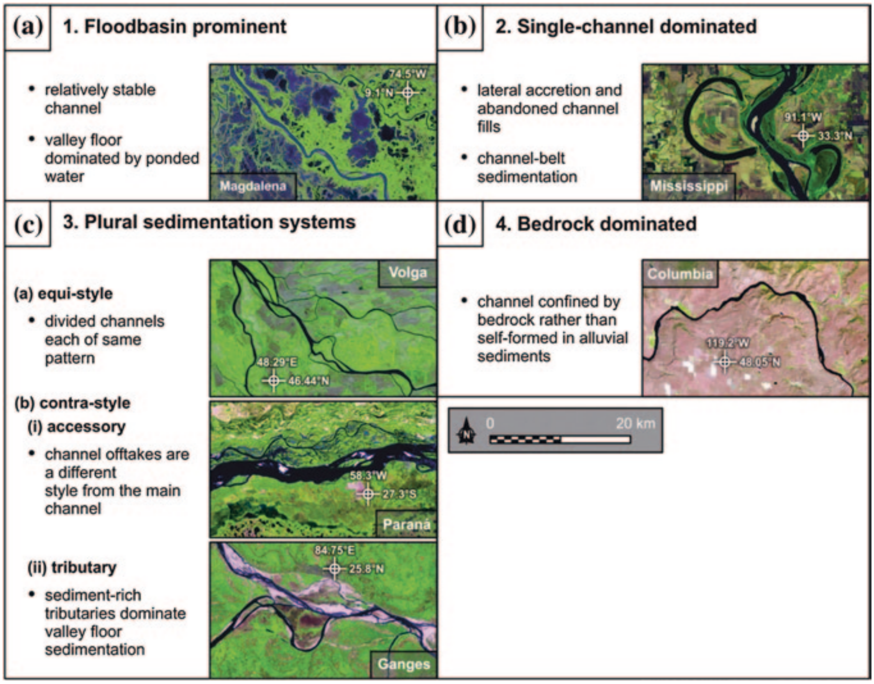


Fig. 7.5 The four main fluvial patterns of large-scale fluvial depositional systems: **a** lacustrine-dominated; **b** dominated by single large river; **c** Large river with many tributaries exhibiting various styles; **d** confined or bedrock-dominated river. Ashworth and Lewin (2012, Fig. 8, p. 94)

followed by the more localized heating and uplift of the rift shoulder of the newly formed Atlantic rift system (Fig. 7.6a).

During the Cenozoic, available evidence suggests that big rivers in the North American continental interior were flowing the other way, towards the east (Fig. 7.6b; McMillan 1973; Duk-Rodkin and Hughes 1994). The reversal in regional drainage is attributed to the formation and uplift of the Cordilleran orogen, with a likely contribution, also, from dynamic topography effects related to mantle thermal systems that were re-ordered as Pangea broke up and the North American continent began to be carried westward.

7.4 Hawkesbury Sandstone: Big River Prediction from Facies Analysis

The challenge in the examination of the rock record is to identify key indicators of scale. As an example of an interpretation of a "big-river" deposit based on facies criteria, this section provides a modified version of a discussion provided in

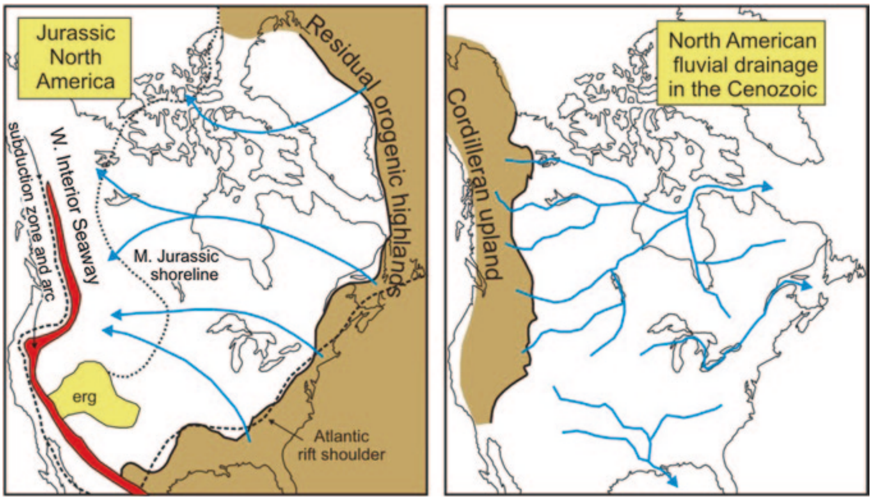


Fig. 7.6 Reconstructions of large river systems in North America, during **a** the Jurassic (Dickinson and Gehrels 2003) and **b** the Cenozoic (McMillan 1973; Duk-Rodkin and Hughes 1994)

Miall (2006b). The example under discussion has long been considered an ancient analog of the sandy braided Brahmaputra River, but it has been a legitimate question how close the comparison is. Does the interpreted scale of the Hawkesbury rivers compare closely to the Brahmaputra, or were they bigger or smaller?

The Hawkesbury Sandstone is Triassic in age, and was deposited in the Sydney Basin, a narrow foreland basin flanking the New England Fold Belt. Regional paleoflow patterns indicate fluvial currents flowing in a northeasterly direction, and provenance studies by Cowan (1993) confirm that the sandstone was sourced from the craton and Lachlan Fold Belt to the south and west. Transport directions were therefore neither clearly axial nor transverse; a cratonic source for a foreland basin fill is unusual.

Miall and Jones (2003) describe the recording and analysis of a 6-km-long profile, almost 100% exposed, of the Hawkesbury Sandstone along the eastern coastline of Kurnell Peninsula, south of Sydney (Fig. 7.7). A simplified version of this profile, with a vertical exaggeration of $\times 16.7$, is shown in Fig. 7.8. Tracing of major bounding surfaces within this profile outlined fifteen major sandbodies, bounded by surfaces of fifth-order rank (classification of Miall (1988, 1996)), that is, the scale of major channels within a braided system (Table 7.1).

The thicknesses of accretionary architectural elements (lateral-accretion and downstream-accretion units, in the terminology of Miall (1988)) in this long outcrop indicate that channel bars ranged up to about 10 m in height (Fig. 7.9). This height indicates the minimum constructional depth of the channels, although bankfull depth was undoubtedly somewhat greater. Scour depths ranged from 4 to 20 m, based on the identification of scour hollows. Second-order channels in the modern Brahmaputra

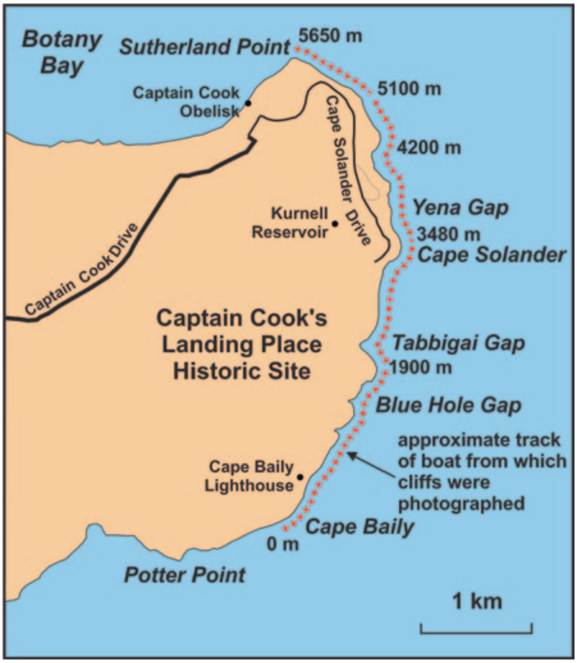


Fig. 7.7 Location of Kurnell Peninsula profile (Fig. 7.8), across Botany Bay immediately to the south of Sydney, Australia

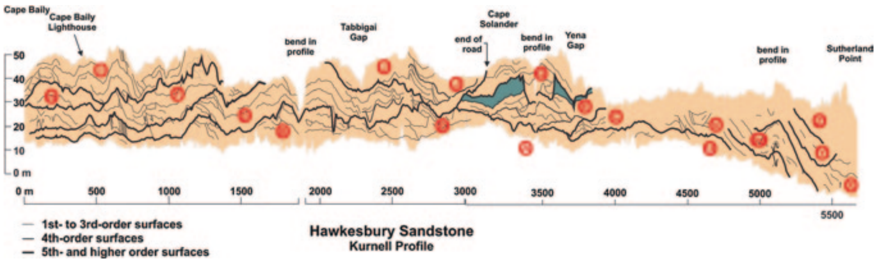


Fig. 7.8 Architectural Profile of the Hawkesbury Sandstone, Kurnell Peninsula, near Sydney, Australia (Location in Fig. 7.7), with ×16.5 vertical exaggeration. The orientation of the profile varies somewhat along the cliff, as indicated by the course of the boat shown in Fig. 7.7. The *top* and *bottom* of the cliff are indicated by the *dotted lines*. Fifth-order architectural elements are labeled with capital letters in circles (from Miall and Jones 2003)

are typically 10–12 m deep, but scour depths of up to 44 m have been recorded (Best and Ashworth 1997). Maximum (bankfull) depths are reached for short periods during each monsoon flood. Bars typically range from half to slightly less than bankfull depth, therefore in a 12-m-deep channel, bars would typically be about 7 m high (Bristow, pers. comm., 2001). Comparisons with the scale of equivalent elements of the modern Brahmaputra river are summarized in Figs. 7.10 and 7.11.

Table 7.1 Size and orientation of the major (fifth-order) elements in the Kurnell Peninsula profile

Element	Width (m)	Max. thickness (m)	Paleocurrent Mean	n	Interpreted outcrop orientation
A	>600?	>8	257	10	Strike
B	>3200?	>20			?
C	>500	>8	297	5	Strike
D	~2700	20	082	3	Strike
E	>900	22	360	9	Dip
F	>1600	>20			?
G	>800	>18	097	12	Strike
H	~600	11			?
I	>1300	>20			?
J	>1100	>18	112	14	Strike
U	?	>10			?
V	>1200	10	135	6	Dip
W	>600	13	342	15	Dip
X	>500	15	101	7	Oblique
Y	>300	11	095	5	Oblique

From Miall and Jones (2003)

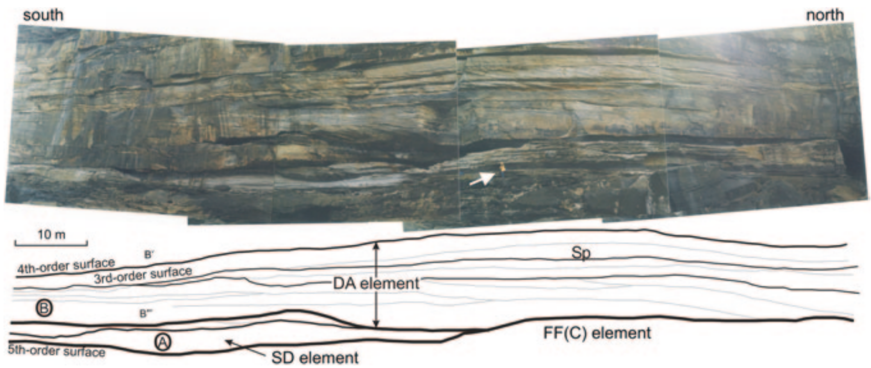


Fig. 7.9 Panorama of a downstream-accretion element (DA), near Curracurrong, Royal National Parks, south of Sydney, Australia. The element rests on a fifth-order bounding surface, and below this is a finer grained sandstone bed, which accounts for the erosional overhang of the base of the DA unit. The element is capped by a fourth-order surface. The scale of this outcrop is indicated by the person, at centre (*arrow*), but the photomosaic was constructed from four photographs all taken from the same point, so the scale is reduced at either end of the view (from Miall and Jones 2003)

Detailed data are now being obtained regarding the internal architecture of modern channel and bar deposits in the Brahmaputra (Jamuna) River. Best et al. (2003) documented the structure and evolutionary history of a mid-channel braid bar, which was initially 1.5 km long, oriented in a downstream direction. Within

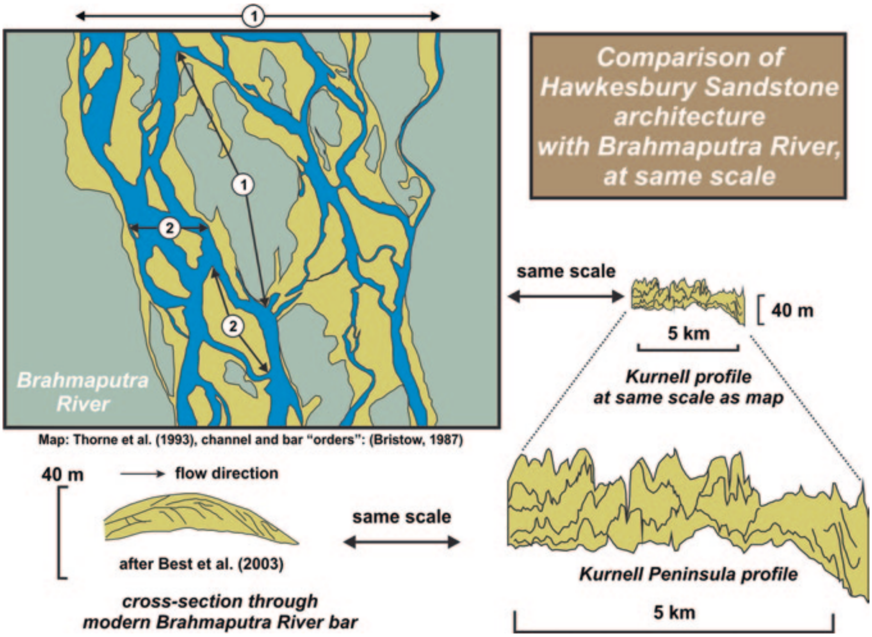


Fig. 7.10 Comparison of the architecture of the Hawkesbury Sandstone at Kurnell Peninsula with the dimensions of channels and bars of the modern Brahmaputra River. At top, a typical reach of the Brahmaputra (Jamuna) River of northern Bangladesh is shown next to a summary of the Kurnell Peninsula outcrop profile reduced to the same horizontal scale. Numbers 1 and 2 in circles indicate the first- and second-order channels and within-channel bar complexes of the river, as classified by Bristow (1987). Note the scale of the interpreted crevasse channel in the Kurnell Peninsula profile. At *bottom*, the Kurnell Peninsula profile is shown at the same scale as a longitudinal cross-section through a modern bar in the Brahmaputra River, as reconstructed from GPR data (from Miall and Jones 2003)

two years the bar had migrated downstream a distance equivalent to its own length and had doubled in size. A simplified version of a long-axis cross-section through this bar, derived from GPR analysis, is shown in Fig. 7.10.

The major problem in estimating river scale from ancient deposits is the ephemeral nature of the modern deposits with which we would make the comparisons. Although we now know more about the internal architecture of a Brahmaputra bar than ever before, we have no way of knowing how much of this deposit will survive long enough to become part of the geological record, nor how typical such a series of depositional and preservational events would be in a river of this type. Miall and Jones (2003) worked through arguments indicating that the preserved deposit widths of a Brahmaputra-scale channel and bar complex could range anywhere from 1 to 6 km.

The largest Hawkesbury Sandstone sand body for which the width is reasonably certain is ~2.7 km wide, in the Kurnell profile. The extremely limited paleocurrent data available from the crossbedding in this body suggests that the

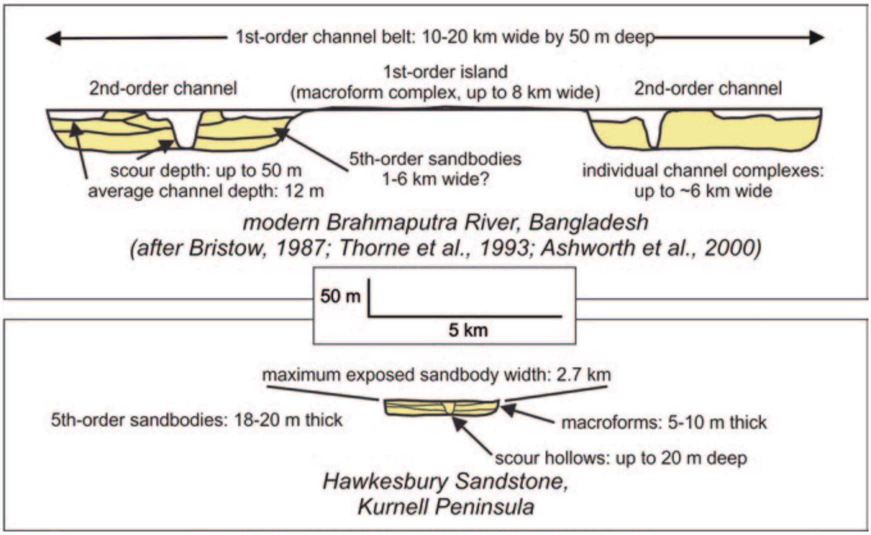


Fig. 7.11 Comparison of sandbody scales between the modern Brahmaputra River and the Hawkesbury Sandstone at the Kurnell Peninsula. The Brahmaputra architecture reconstruction is speculative, based on data provided by Bristow (1987), Thorne et al. (1993), Ashworth et al. (2000) and Best et al. (2003) (from Miall and Jones (2003))

Kurnell Profile is oriented approximately perpendicular to the flow direction within this channel, so that the 2.7 km figure is an approximate measure of sand body width. This is in the middle of the range of possible widths of a typical 2nd-order Brahmaputra sand body (Fig. 7.10). Other Hawkesbury Sandstone bodies appear to be somewhat smaller, although most are incomplete.

Two main conclusions may be drawn from this exercise: (1) Comparisons of the widths and lengths of channels and bars of modern rivers with the scales of similar features in ancient deposits are of very limited use, firstly because it is so difficult to measure or estimate the scales of these features in ancient rocks from outcrop or well data, and secondly, because the preservability of these features in modern rivers is largely unknowable. Seismic time-slice analysis offers the possibility to revolutionize such comparisons, because of the ability to see entire ancient river systems as preserved snapshots (Figs. 4.25, 4.26, 4.30, 4.31, and see example, below). (2) Comparisons of vertical dimensions of channels, bars, scour hollows and bedforms between modern and ancient deposits are likely to be much more reliable, because these features are much more easily measured (in vertical section) and because it is quite easy to assess how typical such measurements are relative to a given river system, either modern or ancient. The vertical dimensions of channels (depths) can be assessed from the heights of cutbanks. The vertical dimensions of bars can be assessed from the amplitude of dipping accretionary bodies. As noted above, care must be taken to ensure that the measurements relate to channel and bar features, not to regional units formed by allogenic processes.

Extreme care needs to be taken when assessing vertical dimensions from drill hole data (logs, core) alone, because of the difficulty in determining the significance (rank) of bounding surfaces within sand bodies. Figure 2.17 and the discussion of this figure in Chap. 2 highlights this problem.

7.5 Plio-Pleistocene Deposits of the Malay Basin: Big River Prediction from Tectonic Setting

The Malaya Basin is one of many basins that developed as part of the process of "extrusion tectonics" that took place throughout southeast Asia following the collision of India with the Asian continent in the early Cenozoic. These basins are mainly bounded by faults on which extensional dip-slip and strike-slip are predominant (Tapponnier et al. 1986; Hutchison 1989). Subsidence and sedimentation have been rapid, with vast quantities of clastic debris shed from rising mountain belts, such as the Himalayas, and transported basinward by such giant rivers as the Brahmaputra, Irrawaddy, Red, and Mekong. Nonmarine sections at least 8 km thick have been reported in the subsurface beneath the Gulf of Thailand (e.g., Pattani Trough: Blanche 1990). Hutchison (1989, p. 69) and Brookfield (1998) suggested that the Mekong River flowed through Thailand and out into the adjacent Gulf until it was diverted by faulting during the late Cenozoic (Fig. 7.12). Much of the detritus was trapped in terminal fluvial and lacustrine basins within Thailand during the Oligocene-Pliocene (O'Leary and Hill 1989; Bidston and Daniels 1992).

The present marine environment of the Gulf of Thailand reflects a rise in relative sea level during the Holocene, as a result of eustatic or tectonic processes, or both, and probably also reflect s a recent reduction in sediment supply, as a result of river diversion. The Mekong River now enters the sea through Vietnam, and the major river of Thailand, the Chao Phraya, drains only the interior of Thailand, and is not classed as one of the world's major rivers. However, the structural setting and tectonic history of the Malay Basin leads to a reasonable prediction that the deposits of a big river might be present in the late Cenozoic section. Seismic data through portions of the Malay basin Plio-Pleistocene section were analyzed by Miall (2002) and Reijenstein et al. (2011) and some of the highlights of those analyses are introduced here in order to illustrate the problems of paleogeographic prediction in a tectonically active area.

None of the rivers identified on this analysis is particularly large. What characterizes them is their variability in fluvial style, which is interpreted as a result of continual tectonic disturbance, high-frequency (Milankovitch) climate change or glacioeustatic sea-level change (or a combination of these processes), modifying the slope, load and discharge conditions on a time scale of 10^3–10^4 year.

Seismic data demonstrate that the Plio-Pleistocene section consists of a series of sequences 5–30 m thick, with incised valleys at their base up to 3 km wide and up to 20 m deep at their base. Figure 6.39 is a cross section through these deposits, and Fig. 4.43 illustrates a series of time slices through one of the

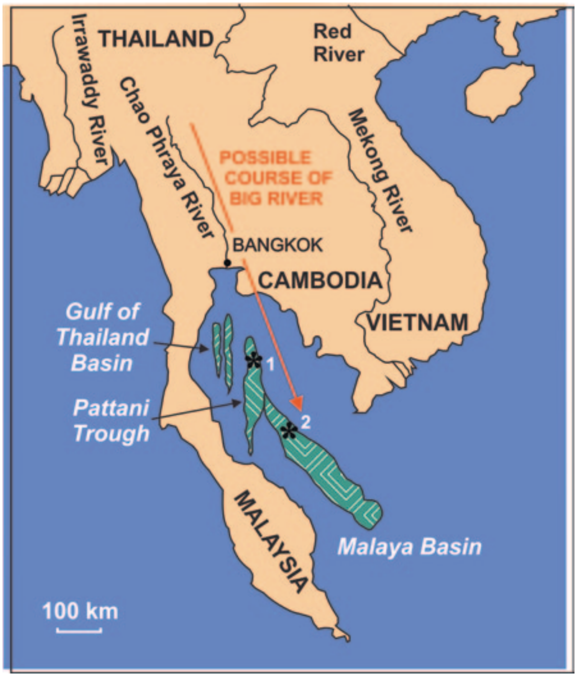


Fig. 7.12 Location of Plio-Pleistocene deposits of the Malay Basin, down-dip from the Chao Phraya River and on the trend of a possible tectonically controlled big river. Seismic images from two projects are discussed in this chapter: 1: Reijenstein et al. (2011); 2: Miall (2002)

valleys, illustrating the architecture of the fill. A meander belt occupies the base of the valley, which is about 10 km wide, and is overlain by a seismically structureless section, probably floodplain deposits, suggesting that the river has been diverted, captured or abandoned. V-shaped tributary valleys can be seen entering from the side. Figure 7.13 illustrates several of the meandering systems imaged by Reijenstein et al. (2011). Figure 7.14 illustrates a time slice and a cross-section through a braided system which has a channel-belt width of about 4 km. This is equivalent in scale to one of the major channels and component bars of the sand-bed Brahmaputra system, but is less than half the width of the entire Brahmaputra channel belt. A closer comparison, in terms of scale, is with the Red River of Oklahoma, which is at the bottom of the scale of "big rivers," using the cut-off limit of 1,000-km length suggested earlier in this chapter.

These are the largest fluvial systems identified within the project area. They suggest the development of rivers of moderately large scale, but not comparable to the present-day trunk drainages in southeast Asia, such as the Irrawaddy (Myanmar), Mekong or Red River (Vietnam).

Did the seismic surveys by Miall (2002) and Reijenstein et al. (2011) miss the largest river systems in this area? This seems unlikely, The purpose of the seismic surveys was to explore the thickest section in the centre of the Malay Basin, which is, itself, one of the largest of the Cenozoic basins that resulted from the Himalayan orogeny in the Gulf of Thailand. It seems more likely that Hutchison's (1989) suggestion is correct, that the major drainage into Thailand and the

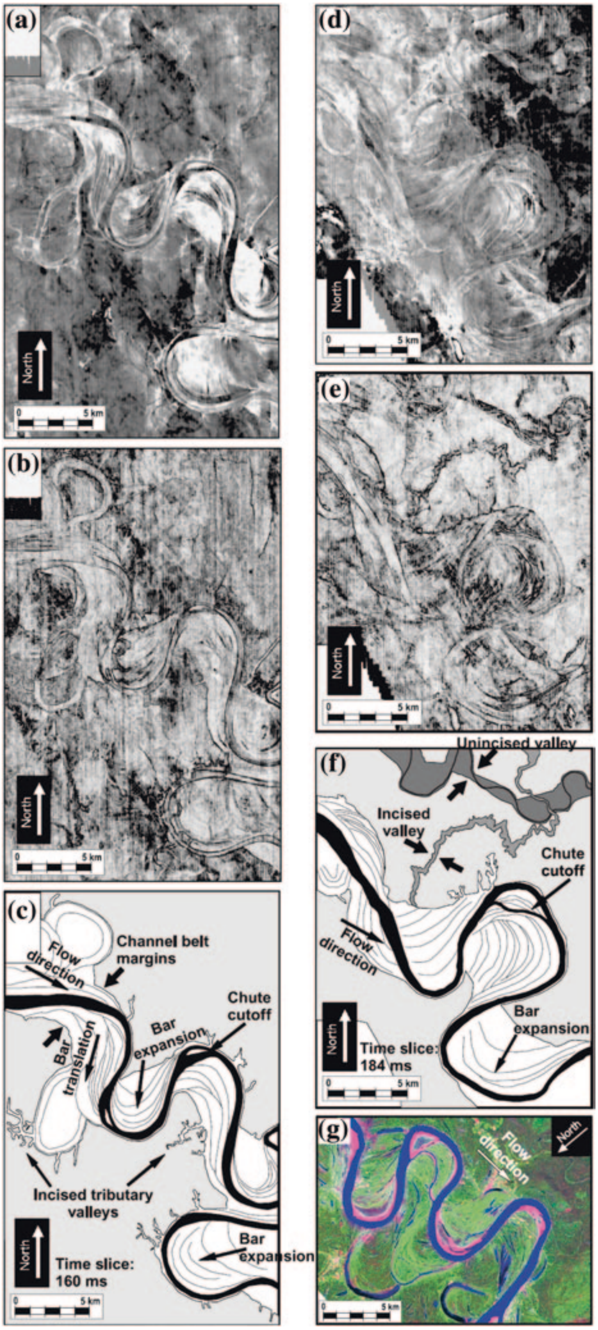


Fig. 7.13 The plan-view images of continental-scale channel and channel belts evident at 160 ms (**a**,**b**, **c**) and 184 ms (**d**, **e**, **f**). Panels **a**, **b**, and **c** represent the 160-ms time slice visualized as **a** conventional amplitude display, **b** coherence seismic attribute, and **c** mapped surface delineating the major geomorphic and depositional elements. Panels **d**, **e**, and **f** represent the 184-ms time slice visualized as **d** conventional amplitude display, **e** coherence seismic attribute, and **f** mapped surface delineating the major geomorphic and depositional elements. Panel G is a modern analog with similar dimensions and comparable depositional elements, Ucayali River, Peru (diagram and caption from Reijenstein et al. 2011, Fig. 6, p. 1970). AAPG © 2011. Reprinted by permission of the AAPG whose permission is required for further use

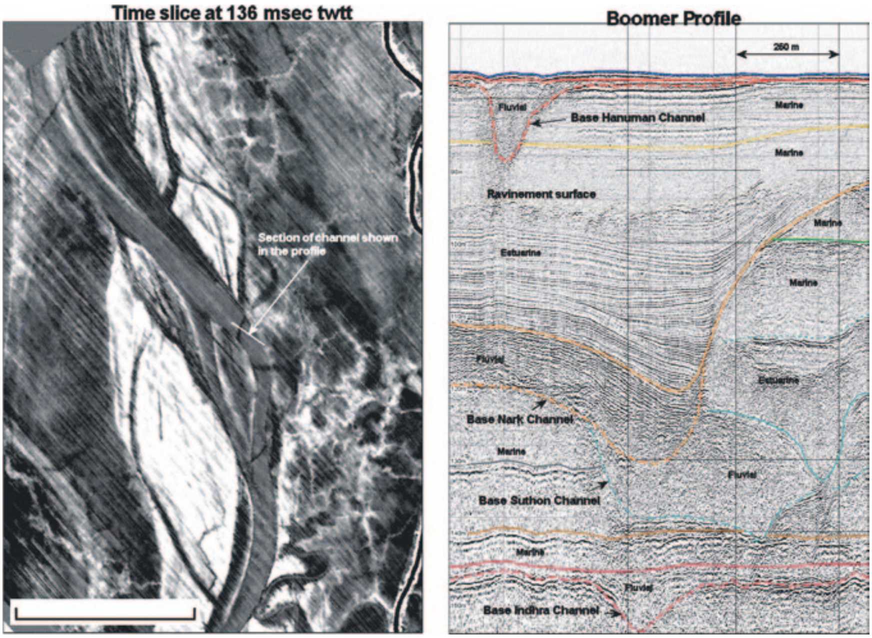


Fig. 7.14 Braided channel system in Plio-Pleistocene deposits of the Malay basin. The image on the *left* corresponds to the "Nauk" channel in the cross-section (Miall 2002)

offshore gulf was diverted into the Mekong prior to the deposition of the fluvial systems illustrated here. Brookfield (1993, 1998) has demonstrated that this type of diversion, resulting in major changes in sediment dispersal paths, has been a common occurrence throughout the Himalayan orogeny. Tectonic setting is, therefore, no reliable predictor of fluvial scale.

7.6 Discussion

The best clue to the scale of a fluvial depositional system is the size of the depositional elements that can be observed, especially where these may be documented in large surface outcrops. It was the enormous scale of the point bars in the Athabasca Oil Sands, Alberta, and the scale of the crossbedding in the Hawkesbury Sandstone, Australia, and in the Neoproterozoic deposits of NW Canada that first alerted sedimentologists to the likely presence of very large river systems in these cases. Scale estimates from vertical profile data, alone, are inadequate, because of the ambiguity of the vertical dimension with respect to autogenic versus allogenic controls. Big rivers may not therefore, be interpreted based on limited subsurface-well data. Seismic data, especially time-slice seismic sections, however, will be definitive.

Tectonic setting is no predictor of fluvial scale. A few specific tectonic settings (foreland basins, strike-slip basins and rift basins) are more likely than others to be characterized by major axial drainage systems because of the funneling effect of the basin configuration, but this is not a useful predictor of overall scale. The largest rivers, as Potter (1978) first showed, are those that flow across major tectonic boundaries within large continents. Their presence is impossible to predict with any certainty, but clues from regional geology may be instructive. Thus, Dickinson's (1988) question about the sources of the large Permian and Jurassic erg systems of the American southwest, which stimulated the search for large, contemporaneous highlands and regional slopes that could have delivered the appropriate amount of sediment (Dickinson and Gehrels 2003). Likewise, a comparable problem stimulated Rainbird et al.'s (1997) analysis of Precambrian sediment sources. A more complex body of evidence was used by Duk-Rodkin and Hughes (1994) to reconstruct the pre-glacial Cenozoic drainage of the cratonic interior of Canada, building on clues from landforms and stratigraphy, some of which were first observed in the nineteenth century. The ultimate fate of much of the detritus eroded from the Cordilleran orogen and craton of central Canada is thought to be the continental margin off Baffin Island and Labrador, but there is, as yet, little confirmatory evidence for this beyond seismic evidence of the scale of these subsea deposits. Drilling data are not yet available.

None of these regional studies have yet pinpointed actual large rivers. In each case, the main fluvial channels, large though they might have been, are probably entirely lost to erosion, although detailed mapping of buried cratonic unconformities might prove fruitful. Potentially, the Permian/Jurassic valley systems crossing the cratonic interior of North America could be preserved. This interval corresponds to the Absaroka sequence of Sloss (1963), which is represented by a regional unconformity across much of the High Plains region of Canada and adjacent areas of the northern United States.

References

Adams MM, Bhattacharya JP (2005) No change in fluvial style across a sequence boundary, Cretaceous Blackhawk and Castlegate formations of central Utah, USA. J Sediment Res 75:1038–1051

Ager DV (1973) The nature of the stratigraphical record. John Wiley, New York 114 p

Aigner T, Asprion U, Hornung J, Junghans W-D, Kostrewa R (1996) Integrated outcrop analogue studies for Triassic alluvial reservoirs: examples from southern Germany. J Petrol Geol 19:393–406

Akinpelu OC (2010) The use of ground-penetrating radar (GPR) to explore the architecture of nonmarine clastic deposits: unpublished Ph.D. thesis, University of Toronto

Alexander J (1992) A discussion of alluvial sandstone body characteristics related to variations in marine influence, Middle Jurassic of the Cleveland Basin, UK, and the implications for analogous Brent Group strata in the North Sea Basin. Geol Soc London (Special Publication) 61:149–167

Allen JP, Fielding CR (2007) Sedimentology and stratigraphic architecture of the Late Permian Betts Creek Beds, Queensland, Australia. Sediment Geol 202:5–34

Allen JP, Fielding CR, Gibling MR, Rygel MC (2011) Fluvial response to paleo-equatorial climate fluctuations during the late Paleozoic ice age. Geol Soc Am Bull 123:1524–1538

Allen JRL (1963a) Henry Clifton Sorby and the sedimentary structures of sands and sandstones in relation to flow conditions. Geologie en Mijnbouw 42:223–228

Allen JRL (1963b) Depositional features of Dittonian rocks: Pembrokeshire compared with the Welsh Borderlands. Geol Mag 100:385–400

Allen JRL (1964) Studies in fluviatile sedimentation: six cyclothems from the Lower Old Red Sandstone, Anglo-Welsh basin. Sedimentology 3:163–198

Allen JRL (1965a) A review of the origin and characteristics of recent alluvial sediments. Sedimentology 5:89–191

Allen JRL (1965b) The sedimentation and palaeogeography of the Old Red Sandstone of Anglesey, North Wales. Yorkshire Geol Soc Proc 35:139–185

Allen JRL (1966) On bed forms and paleocurrents. Sedimentology 6:153–190

Allen JRL (1974) Studies in fluviatile sedimentation: implications of pedogenic carbonate units, Lower Old Red Sandstone, Anglo-Welsh outcrop. Geol J 9:181–208

Allen JRL (1978) Studies in fluviatile sedimentation: an exploratory quantitative model for the architecture of avulsion-controlled suites. Sediment Geol 21:129–147

Allen JRL (1979) Studies in fluviatile sedimentation: an elementary geometrical model for the connectedness of avulsion-related channel sand bodies. Sediment Geol 24:253–267

Allen JRL (1983) Studies in fluviatile sedimentation: bars, bar complexes and sandstone sheets (low-sinuosity braided streams) in the Brownstones (L. Devonian), Welsh Borders. Sediment Geol 33:237–293

A. Miall, *Fluvial Depositional Systems*, Springer Geology,
DOI: 10.1007/978-3-319-00666-6,

Allen PA (2008) Time scales of tectonic landscapes and their sediment routing systems. In: Gallagher K, Jones SJ, Wainwright J (eds) Landscape evolution: denudation, climate and tectonics over different time and space scales, vol. 296. Geological Society, London, Special Publication, pp 7–28

Allen PA, Collinson JD (1986) Lakes. In: Reading HG (ed) Sedimentary environments and facies. Blackwell Scientific Publications, Oxford, pp 63–94

Amarosi A, Pavesi M, Ricci Lucchi M, Sarti G, Piccin A (2008) Climatic signature of cyclic fluvial architecture from the Quaternary of the central Po Plain, Italy. Sediment Geol 209:58–68

Anadón P, Cabrera L, Colombo F, Marzo M, Riba O (1986) Syntectonic intraformational unconformities in alluvial fan deposits, eastern Ebro basin margins (NE Spain). In: Allen PA, Homewood P (eds) Foreland basins: international association of sedimentologists, vol. 8. Special Publication, pp 259–271

Andersen T, Zachariassen E, Høye T, Meisingset HC, Otterlei C, van Wijngaarden AJ, Hatland K, Mangerøy G, Liestol FM (2006) Method for conditioning the reservoir model on 3D and 4D elastic inversion data applied to a fluvial reservoir in the North Sea. Society of Petroleum Engineers, paper SPE 100190

Antia J, Fielding CR (2011) Sequence stratigraphy of a condensed low-accommodation succession: lower upper Cretaceous Dakota Sandstone, henry Mountains, Southeastern Utah. Am Assoc Petr Geol B 95:413–447

Archer AW, Greb SF (1995) An Amazon-scale drainage system in the Early Pennsylvanian of central North America. J Geol 103:611–628

Archer SG, Reynisson RF, Schwab AM (2011) River terraces in thew rock record: an overlooked landform in geological interpretation. In: Davidson SK, Leleu S, North CP (eds) From river to rock record: society for sedimentary geology (SEPM), vol. 97. Special Publication, pp 63–85

Ardies GW, Dalrymple RW, Zaitlin BA (2002) Controls on the geometry of incised valleys in the Basal Quartz unit (Lower Cretaceous), Western Canada Sedimentary Basin. J Sediment Res 72:602–618

Arnott RWC, Zaitlin BA, Potocki DJ (2002) Stratigraphic response to sedimentation in a newt-accommodation-limited setting, Lower Cretaceous basal Quartz, south-central Alberta. B Can Petrol Geol 50:92–104

Aschoff JL, Steel RJ (2011) Anomalous clastic wedge development during the Sevier-Laramide transition, North American Cordilleran foreland basin, USA. Geol Soc Am Bull 123:1822–1835

Ashley GM (1990) Classification of large-scale subaqueous bedforms: a new look at an old problem. J Sediment Petrol 60:160–172

Ashworth PJ, Best JL, Peakall J, Lorsong JA (1999) The influence of aggradation rate on braided alluvial architecture: field study and physical scale-modelling of the Ashburton River gravels, Canterbury Plains, New Zealand. In: Smith ND, Rogers J (eds) Fluvial sedimentology VI, vol. 28. Special Publication. International Association of Sedimentologists, pp 333–346

Ashworth PJ, Best JL, Jones M (2004) Relationship between sediment supply and avulsion frequency in braided rivers. Geology 32:21–24

Ashworth PJ, Best JL, Jones M (2007) The relationship between channel avulsion, flow occupancy and aggradation in braided rivers: insights from an experimental model. Sedimentology 54:497–513

Ashworth PJ, Best JL, Roden JE, Bristow CS, Klaassen GJ (2000) Morphological evolution and dynamics of a large, sand braid-bar, Jamuna river, Bangladesh. Sedimentology 47:533–555

Ashworth PJ, Lewin J (2012) How do big rivers come to be different? Earth Sci Rev 114:84–107

Aslan A, Autin WJ (1999) Evolution of the Holocene Mississippi River floodplain, Ferriday, Louisiana: insights of the origin of fine-grained floodplains. J Sediment Res 69:800–815

Aslan A, Autin WJ, Blum MD (2005) Causes of river avulsion: insights from the Late Holocene avulsion history of the Mississippi River. J Sediment Res 75:650–664

Aslan A, White WA, Warne AG, Guevara EH (2003) Holocene evolution of the western Orinoco Delta, Venezuela. Geol Soc Am Bull 115:479–498

Asprion U, Aigner T (1999) Towards realistic aquifer models: three-dimensional georadar surveys of Quaternary gravel deltas (Singen Basin, SW Germany). Sediment Geol 129:281–297

Atchley SC, Nordt LC, Dworkin SL (2004) Eustatic control on alluvial sequence stratigraphy: a possible example from the Cretaceous-tertiary transition of the Tornillo Basin, Big Bend National Park, West Texas, USA. J Sediment Res 74:391–404

Autin WJ, Burns SF, Miller BJ, Saucier RT, Snead J (1991) Quaternary geology of the lower Mississippi valley. In: Morrison RB (ed) Quaternary nonglacial geology: conterminous US, vol. K-2. Geological Society of America, Geology of North America, Boulder, CO, pp 547–582

Bailey RJ, Smith DG (2010) Scaling in stratigraphic data series: implications for practical stratigraphy. First Break 28:57–66

Barrier L, Proust J-N, Nalpas T, Robin C, Guillocheau F (2010) Control of alluvial sedimentation at foreland-basin active margins: a case study from the northeastern Ebro Basin (southeastern Pyrenees, Spain). J Sediment Res 80:728–749

Batson PA, Gibling MR (2002) Architecture of channel bodies and paleovalley fills in high-frequency Carboniferous sequences, Sydney Basin, Atlantic Canada. Bull Can Petrol Geol 50:138–157

Bellman LW (2010) Integrated 3D reservoir characterization for oil sands evaluation, development and monitoring. Search and Discovery Article #40541, American Association of Petroleum Geologists

Beres M, Huggenberger P, Green AG, Horstmeyer H (1999) Using two- and three-dimensional georadar methods to characterize glaciofluvial architecture. Sediment Geol 129:1–24

Best JL, Ashworth PJ (1997) Scour in large braided rivers and the recognition of sequence stratigraphic boundaries. Nature 387:275–277

Best JL, Ashworth PJ, Bristow CS, Roden J (2003) Three-dimensional sedimentary architecture of a large mid-channel sand braid bar, Jamuna River, Bangladesh. J Sediment Res 73:516–530

Berendsen HJA, Stouthamer E (2001) Palaeogeographic development of the Rhine-Meuse delta. Koninklijke Van Gorcum, Assen, The Netherlands, p 268

Beutner EC, Flueckinger LA, Gard TM (1967) Bedding geometry in a Pennsylvanian channel sandstone. Geol Soc Am Bull 78:911–916

Bhattacharya JP (1993) The expression and interpretation of marine flooding surfaces and erosional surfaces in core: examples from the Upper Cretaceous Dunvegan Formation in the Alberta foreland basin. In: Summerhayes CP, Posamentier HW (eds) Sequence stratigraphy and facies associations, International Association of Sedimentologists, vol. 18. Special Publication, pp 125–160

Bhattacharya JP (2011) Practical problems in the application of the sequence stratigraphic method and key surfaces: integrating observations from ancient fluvial-deltaic wedges with Quaternary and modeling studies. Sedimentology 58:120–169

Bidston BJ, Daniels JS (1992) Oil from the ancient lakes of Thailand. In: National conference on geologic resources of Thailand: potential for future development. Department of Mineral Resources, Bangkok, Thailand, pp 584–599

Blair TC, Bilodeau WL (1988) Development of tectonic cyclothems in rift, pull-apart, and foreland basins: sedimentary response to episodic tectonism. Geology 16:517–520

Blanche JB (1990) An overview of the exploration history and hydrocarbon potential of Cambodia and Laos. SEAPEX Proc 9:89–99

Blum MD (1993) Genesis and architecture of incised valley fill sequences: a late Quaternary example from the Colorado River, Gulf Coastal Plain of Texas. In: Weimer P, Posamentier HW (eds) Siliciclastic sequence stratigraphy: recent developments and applications, vol. 58. American Association of Petroleum Geologists Memoir, pp 259–283

Blum M, Martin J, Milliken K, Garvin M (2013) Paleovalley systems: insights from Quaternary analogs and experiments. Earth Sci Rev 116:128–169

Blum MD, Törnqvist TE (2000) Fluvial responses to climate and sea-level change: a review and look forward. Sedimentology 47:2–48

Bridge JS (1993) The interaction between channel geometry, water flow, sediment transport and deposition in braided rivers. In: Best JL, Bristow CS (eds) Braided rivers, vol. 75. Geological Society, London, Special Publication, pp 13–71

Bridge JS (1985) Paleochannel patterns inferred from alluvial deposits: a critical evaluation. J Sediment Petrol 55:579–589

Bridge JS (1993) Description and interpretation of fluvial deposits: a critical perspective. Sedimentology 40:801–810

Bridge JS (2003) Rivers and floodplains: forms, processes and sedimentary record. Blackwell, Oxford, 491 p

Bridge JS, Collier R, Alexander J (1998) Large-scale structure of Calamus River deposits (Nebraska, USA) revealed using ground penetrating radar. Sedimentology 45:977–986

Bridge JS, Leeder MR (1979) A simulation model of alluvial stratigraphy. Sedimentology 26:617–644

Bridge JS, Mackey SD (1993a) A revised alluvial stratigraphy model. In: Marzo M, Puigdefábregas C (eds) Alluvial sedimentation, vol. 17. International Association of Sedimentologists, Special Publication, pp 319–336

Bridge JS, Mackey SD (1993b) A theoretical study of fluvial sandstone body dimensions. In: Flint SS, Bryant ID (eds) The geological modelling of hydrocarbon reservoirs and outcrop analogues, vol. 15. International Association of Sedimentologists, Special Publication, pp 213–236

Bridge JS, Mackey SD (1993c) A theoretical study of fluvial sandstone body dimensions. In: Flint SS, Bryant ID (eds) The geological modelling of hydrocarbon reservoirs and outcrop analogues, vol. 15. International Association of Sedimentologists, Special Publication, pp 213–236

Bridge JS, Tye RS (2000) interpreting the dimensions of ancient fluvial channel bars, channels, and channel belts from wire-line logs and cores. Am Assoc Petr Geol B 84:1205–1228

Brierley GJ (1989) River planform facies models: the sedimentology of braided, wandering and meandering reaches of the Squamish River. Sediment Geol 61:17–36

Brierley GJ (1991a) Bar sedimentology of the Squamish River, British Columbia: definition and application of morphostratigraphic units. J Sediment Petrol 61:211–225

Brierley GJ (1991b) Floodplain sedimentology of the Squamish River, British Columbia: relevance of element analysis. Sedimentology 38:735–750

Brierley GJ (1996). Channel morphology and element assemblages: a constructivist approach to facies modeling. In: Carling PA, Dawson MR (eds) Advances in fluvial dynamics and stratigraphy.John Wiley and Sons, Chichester, pp 263–298

Brierley GJ, Hickin EJ (1991) Channel planform as a non-controlling factor in fluvial sedimentology: the case of the Squamish River floodplain, British Columbia. Sediment Geol 75:67–83

Bristow CS (1999) Gradual avulsion, river metamorphosis and reworking by underfit streams: a modern example from the Brahmaputra River in Bangladesh and a possible ancient example in the Spanish Pyrenees. In Smith ND, Rogers J (eds) Fluvial sedimentology VI, vol. 28. International Association of Sedimentologists, Special Publication, pp 221–230

Bristow CS (ed) (2003) Ground penetrating radar in sediments, vol. 211. Geological Society, London, Special Publication, 330 p

Bristow CS, Jol HM (2003) An introduction to ground penetrating radar in sediments. In: Bristow CS (ed) Ground penetrating radar in sediments, vol. 211. Geological Society, London, Special Publication, pp 1–7

Brizga SO, Finlayson BL (1990) Channel avulsion and river metamorphosis: The case of the Thomson River, Victoria, Australia. Earth Surf Proc Land 15:391–404

Brown AR (2011) Interpretation of three-dimensional seismic data, 7th edn., vol. 42. American Association of Petroleum Geologists Memoir, 646 p

Browne GH, Naish TR (2003) Facies development and sequence architecture of a late Quaternary fluvial-marine transition, Canterbury Plains and shelf, New Zealand: implications for forced regressive deposits. Sediment Geol 158:57–86

Bryant ID, Flint SS (1993) Quantitative clastic reservoir geological modelling: problems and perspectives. In: Flint SS, Bryant ID (eds) The geological modelling of hydrocarbon reservoirs and outcrop analogues, vol. 15. International Association of Sedimentologists, Special Publication, pp 3–20

Bryant M, Falk P, Paola C (1995) Experimental study of avulsion frequency and rate of deposition. Geology 23:365–369

Burbank DW, Meigs A, Brozovic N (1996) Interactions of growing folds and coeval depositional systems. Basin Res 8:199–223

Butcher SW (1990) The nickpoint concept and its implications regarding on lap to the stratigraphic record. In: Cross TA (ed) Quantitative dynamic stratigraphy. Prentice-Hall, Englewood Cliffs, pp 375–385

Carling PA, Dawson MR (eds) (1996) Advances in fluvial dynamics and stratigraphy. John Wiley and Sons, Chichester, 530 p

Carrigy MA (1971) Deltaic sedimentation in Athabasca Tar Sands. Am Assoc Petr Geol Bull 55:1155–1169

Carson MA (1984) The meandering-braided river threshold: a reappraisal. J Hydrol 73:315–334

Catuneanu O (2006) Principles of sequence stratigraphy. Elsevier, Amsterdam, 375 p

Cecil CB (1990) Paleoclimate controls on stratigraphic repetition of chemical and siliciclastic rocks. Geology 18:533–536

Chatanantavet P, Lamb MP, Nittrouer JA (2012) Backwater controls of avulsion location on deltas. Geophys Res Lett 39(L01402):6

Chawner WD (1935) Alluvial fan flooding, the Montrose, California, flood of 1934. Geogr Rev 25:225–263

Cleveland DM, Atchley SC, Nordt LC (2007) Continental sequence stratigraphy of the upper Triassic (Norian–Rhaetian) Chinle strata, Northern New Mexico, USA: allocyclic and autocyclic origins of paleosol-bearing alluvial successions. J Sediment Res 77:909–924

Cloetingh S (1988) Intraplate stress: a new element in basin analysis. In: Kleinspehn K, Paola C (eds) New perspectives in basin analysis. Springer-Verlag, New York, pp 205–230

Coleman JM (1969) Brahmaputra river: channel processes and sedimentation. Sediment Geol 3:129–239

Coleman JM, Wright LD (1975) Modern river deltas: variability of processes and sand bodies. In: Broussard ML (ed) Deltas, models for exploration. Houston Geological Society, Houston, TX, pp 99–150

Collinson JD (1978) Vertical sequence and sand body shape in alluvial sequences. In: Miall AD (ed) Fluvial sedimentology, vol. 5. Canadian Society of Petroleum Geologists Memoir, pp 577–586

Collinson JD (1986) Alluvial sediments. In: Reading HG (ed) Sedimentary environments and facies. Blackwell Scientific Publications Ltd, Oxford, pp 20–62

Colombera L, Felletti F, Mountney NP, McCaffrey WD (2012) A database approach for constraining stochastic simulations of the sedimentary heterogeneity of fluvial reservoirs. Am Assoc Petr Geol Bull 96:2143–2166

Conaghan PJ, Jones JG (1975) The Hawkesbury Sandstone and the Brahmaputra: a depositional model for continental sheet sandstones. J Geol Soc Australia 22:275–283

Corbeanu RM, McMechan GA, Szerbiak RB, Soegaard K (2002) Prediction of 3-D fluid permeability and mudstone distributions from ground-penetrating radar (GPR) attributes: examples from the Cretaceous Ferron Sandstone Member, east–central Utah. Geophysics 67:1495–1504

Cowan EJ (1991) The large-scale architecture of the fluvial Westwater Canyon Member, Morrison Formation (Jurassic), San Juan Basin, New Mexico. In: Miall AD, Tyler N (eds) The three-dimensional facies architecture of terrigenous clastic sediments, and its implications for hydrocarbon discovery and recovery, vol. 3. Society of Economic Paleontologists and Mineralogists, Concepts in Sedimentology and Paleontology, pp 80–93

Cowan EJ (1993) Longitudinal fluvial drainage patterns within a foreland basin-fill: Permo-Triassic Sydney Basin, Australia. Sediment Geol 85:557–577

Cullingford RA, Davidson DA, Lewin J (1980) Timescales in geomorphology. Wiley, Chichester, 360 p

Currie BS (1997) Sequence stratigraphy of nonmarine Jurassic-Cretaceous rocks, central Cordilleran foreland-basin system. Geol Soc Am Bull 109:1206–1222

Dalrymple RW, Boyd R, Zaitlin BA (eds) (1994) Incised-valley systems: origin and sedimentary sequences, vol. 51. Society of Economic Paleontologists and Mineralogists, Special Publication, 391p

Dalrymple RW, Leckie DA, Tillman RW (eds) (2006) Incised valleys in time and space, vol. 85. Society for Sedimentary Geology (SEPM), Special Publication, 343 p

Davies NS, Gibling MR (2010a) Cambrian to Devonian evolution of alluvial systems: the sedimentological impact of the earliest land plants. Earth-Sci Rev 98:171–200

Davies NS, Gibling MR (2010b) Paleozoic vegetation and the Siluro-Devonian rise of fluvial lateral accretion sets. Geology 38:51–54

Davies NS, Gibling MR (2011) Evolution of fixed-channel alluvial plains in response to Carboniferous vegetation. Nat Geosci 4:629–633

Davies RJ, Posamentier HW, Wood LJ, Cartwright JA (eds) (2007) Seismic geomorphology: applications to hydrocarbon exploration and production, vol. 277. Geological Society, London, Special Publication, 274 p

Díaz-Molina M, Muñoz-García MB (2010) Sedimentary facies and three-dimensional reconstructions of upper Oligocene meander belts from the Loranca Basin, Spain. Am Assoc Petr Geol B 94:241–257

Dickinson WR (1988) Provenance and sediment dispersal in relation to paleotectonics and paleogeography of sedimentary basins. In: Kleinspehn KL, Paola C (eds) New perspectives in basin analysis. Springer Verlag, New York, pp 1–25

Dickinson WR, Gehrels GE (2003) U-Pb ages of detrital zircons from Permian and Jurassic eolian sandstones from the Colorado Plateau, USA: paleogeographic implications. Sediment Geol 163:29–66

Dickinson WR, Soreghan GS, Giles KA (1994) Glacio-eustatic origin of Permo-Carboniferous stratigraphic cycles: evidence from the southern Cordilleran foreland region. In: Dennison JM, Ettensohn FR (eds) Tectonic and eustatic controls on sedimentary cycles, vol. 4. Society for Sedimentary Geology, Concepts in Sedimentology and Paleontology, pp 25–34

Donselaar M, Overeem I (2008) Connectivity of fluvial point-bar deposits: An example from the Miocene Huesca fluvial fan, Ebro Basin, Spain. Am Assoc Petr Geol B 92:1109–1129

Dubiel RF, Hasiotis ST (2011) Deposystems, paleosols, and climatic variability in a continental system: the Upper Triassic Chinle Formation, Colorado Plateau, USA. In: Davidson SK, Leleu S, North CP (eds) From river to rock record, vol. 97. Society for Sedimentary Geology (SEPM), Special Publication, pp 393–421

Dubriel-Boisclair C, Gloaguen E, Marcotte D, Giroux B (2011) Heterogeneous aquifer characterization form ground-penetrating radar tomography and borehole hydrogeophysical data using nonlinear Bayesian simulations. Geophysics 76:J13–J25

Duk-Rodkin A, Hughes OL (1994) Tertiary-Quaternary drainage of the pre-glacial Mackenzie Basin. Quatern Int 22(23):221–241

Eberth DA, Miall AD (1991) Stratigraphy, sedimentology and evolution of a vertebrate-bearing, braided to anastomosed fluvial system, Cutler Formation (Permian-Pennsylvanian), north-central New Mexico. Sediment Geol 72:225–252

Elder WP, Gustason ER, Sageman BB (1994) Correlation of basinal carbonate cycles to near shore para sequences in the Late Cretaceous Greenhorn seaway, Western Interior, USA. Geol Soc Am Bull 106:892–902

Embry AF (1990) A tectonic origin for third-order depositional sequences in extensional basins—implications for basin modeling. In: Cross TA (ed) Quantitative dynamic stratigraphy. Prentice-Hall, Englewood Cliffs, pp 491–501

Ethridge FG (2011) Interpretation of ancient fluvial channel deposits: review and recommendations. In: Davidson SK, Leleu S, North CP (eds) From river to rock record, vol. 97. Society for Sedimentary Geology (SEPM) Special Publication, pp 9–35

Ethridge FG, Schumm SA (1978) Reconstructing paleochannel morphologic and flow characteristics: methodology, limitations and assessment. In: Miall AD (ed) Fluvial sedimentology, vol. 5. Canadian Society of Petroleum Geologists Memoir, pp 703–721

Ethridge FG, Schumm SA (2007) Fluvial seismic geomorphology: a view from the surface. In: Davies RJ, Posamentier HW, Wood,LJ, Cartwright JA (eds) Seismic geomorphology: applications to hydrocarbon exploration and production, vol. 277. Geological Society, London, Special Publication, pp 205–222.

Ettensohn FR (2008) The Appalachian foreland basin in eastern United States. In: Miall AD (ed) The Sedimentary Basins of the United States and Canada: Sedimentary basins of the World, vol 5. Hsü, KJ Series EditorElsevier Science, Amsterdam, pp 105–179

Eugster HP, Hardie LA (1975) Sedimentation in an ancient playa-lake complex: The Wilkins peak member of the Green river formation of Wyoming. Geol Soc Am Bull 86:319–334

Fagan SD, Nanson GC (2004) The morphology and formation of floodplain surface channels, Cooper Creek, Australia. Geomorphology 60:107–126

Fahnestock RK (1963) Morphology and hydrology of a glacial stream—White River, Mt. Rainer, U.S. Geological Survey Professional Paper 422-A, Washington, pp 1–70

Farrell KM (1987) Sedimentology and facies architecture of overbank deposits of the Mississippi River, False River region, Louisiana. In: Ethridge FG, Flores RM, Harvey MD (eds) Recent developments in fluvial sedimentology, vol. 39. Society of Economic Paleontologists and Mineralogists, Special Publication, pp 111–120

Fielding CR (2007) Sedimentology and stratigraphy of large river deposits: recognition in the ancient record, and distinction from incised valley fills. In: Gupta A (ed) Large rivers: geomorphology and management. John Wiley and Sons, Chichester, pp 97–113

Fielding CR, Alexander J, McDonald R (1999) Sedimentary facies from ground-penetrating radar surveys of the modern, upper Burdekin River of north Queensland, Australia: consequences of extreme discharge fluctuations. In: Smith ND, Rogers J (eds) Fluvial sedimentology VI, vol. 28. International Association of Sedimentologists, Special Publication, pp 347–362

Fielding CR, Allen JP, Alexander J, Gibling MR, Rygel MC, Calder JH (2011) Fluvial systems and their deposits in hot, seasonal semiarid and subhumid settings: modern and ancient examples. In: Davidson SK, Leleu S, North CP (eds) From river to rock record, vol. 97. Society for Sedimentary Geology (SEPM), Special Publication, pp 89–111

Fielding CR, Allen JP, Alexander J, Gibling MR (2009) A facies model for fluvial systems in the seasonal tropics and subtropics. Geology 37:623–626

Fielding CR, Ashworth PJ, Best JL, Prokocki EW, Sambrook-Smith GH (2012) Tributive, distributive and other fluvial systems: what really represents the norm in the continental rock record? Sediment Geol 261–262:15–32

Fielding CR, Crane RC (1987) An application of statistical modelling to the prediction of hydrocarbon recovery factors in fluvial reservoir sequences. In: Ethridge FG, Flores RM, Harvey MD (eds) Recent developments in fluvial sedimentology, vol. 39. Society of Economic Paleontologists and Mineralogists, Special Publication, pp 321–327

Fisk HN (1944) Geological investigation of the alluvial valley of the lower Mississippi River. Mississippi River Commission, Vicksburg, Mississippi, 78 p

Fisk HN (1952) Geological investigation of the Atchafalaya Basin and the problem of the Mississippi River diversion. US Army Corps of Engineers, Waterways Experiment Station, Vicksburg, Mississippi, 145 p

Foreman BZ, Heller PL, Clementz MT (2012) Fluvial response to abrupt global warming at the Palaeocene/Eocene boundary. Nature 491:92–95

Friedkin JF (1945) A laboratory study of the meandering of alluvial rivers. Mississippi River Commission, Vicksburg

Friend PF (1983) Towards the field classification of alluvial architecture or sequence. In: Collinson JD, Lewin J (eds) Modern and ancient fluvial systems, vol. 6. International Association of Sedimentologists, Special Publication, pp 345–354

Friend PF, Sinha R (1993) Braiding and meandering parameters. In: Best JL, Bristow CS (eds) Braided rivers, vol. 75. Geological Society, London, Special Publication, pp 105–111

Friend PF, Slater MJ, Williams RC (1979) Vertical and lateral building of river sandstone bodies, Ebro Basin, Spain. J Geol Soc, London 136:39–46

Fustic M, Skulski L, Hanson W, Vanhooren D, Bessette P, Hiunks D, Bellman L, Leckie D (2008) Geological mapping and reservoir characterization of oil sands reservoir by integrating 3D seismic, dipmeter, core descriptions, and analogs in the McMurray Formation, NE Alberta: Search and Discovery Article #40281, American Association of Petroleum Geologists

Galloway WE (1981) Depositional architecture of Cenozoic Gulf Coastal plain fluvial systems. In: Ethridge FG, Flores RM (eds) Recent and ancient nonmarine depositional environments: models for exploration, vol. 31. Society of Economic Paleontologists and Mineralogists Special Publication, pp 127–155

Galloway WE (1989) Genetic stratigraphic sequences in basin analysis II: application to northwest Gulf of Mexico Cenozoic basin. Am Assoc Petr Geol B 73:143–154

Galloway WE (2005) Gulf of Mexico Basin depositional record of Cenozoic North American drainage basin evolution. In: Blum MD, Marriott SB, Leclair SF (eds) Fluvial sedimentology VII, vol. 35, International Association of Sedimentologists, Special Publication, pp 409–423

Gawthorpe RL, Collier REL, Alexander J, Bridge JS, Leeder MR (1993) Ground penetrating radar: application to sand body geometry and heterogeneity studies. In: North CE, Prosser DJ (eds) Characterization of fluvial and aeolian reservoirs, vol. 73. Geological Society, London, Special Publication, pp 421–432

Geehan G (1993) The use of outcrop data and heterogeneity modelling in development planning. In: Eschard R, Doligez B (eds) Subsurface reservoir characterization from outcrop observations. Institut Français du Petrole. ÉditionsTechnip, Paris, pp 53–64

Gibling MR (2006) Width and thickness of fluvial channel bodies and valley fills in the geological record: a literature compilation and classification. J Sediment Res 76:731–770

Gibling MR, Fielding CR, Sinha R (2011) Alluvial valleys and alluvial sequences: towards a geomorphic assessment. In: Davidson SK, Leleu S, North CP (eds) From river to rock record: the preservation of fluvial sediments and their subsequent interpretation, vol. 97. Society for Sedimentary Geology (SEPM), Special Publication, pp 423–447

Godin PD (1991) Fining-upward cycles in the sandy braided-river deposits of the West water Canyon Member (Upper Jurassic), Morrison Formation, New Mexico. Sediment Geol 70:61–82

Gole CV, Chitale SV (1966) Inland delta building activity of Kosi River. J Hydr Div, Proc Am Soc Civil Eng 92(HY2):111–126

Gradzinski R, Baryla J, Doktor M, Gmur D, Gradzinski M, Kedzior A, Paszkowski M, Soja R, Zielinkski T, Zurek S (2003) Vegetation-controlled modern anastamosing system of the upper Narew River (NE Poland) and its sediments. Sediment Geol 157:253–276

Guccione MJ, Burford MF, Kendall JD (1999) Pemiscot Bayou, a large distributary of the Mississippi River and a possible failed avulsion. In: Smith ND, Rogers J (eds) Fluvial sedimentology VI, vol. 28. International Association of Sedimentologists, Special Publication, pp 211–220

Gupta A (2007) Introduction. In: Gupta A (ed) Large rivers: geomorphology and management. John Wiley and Sons, Chichester, pp 1–5

Hajek EA, Heller PL, Sheets BA (2010) Significance of channel-belt clustering in alluvial basins. Geology 38:535–538

Hammon WS III, Zeng X, Corbeanu RM, McMechan GA (2002) Estimation of the spatial distribution of fluid permeability from surface and tomographic GPR data and core, with a 2-D example from the Ferron Sandstone, Utah. Geophysics 67:1505–1515

Hampson GJ, Jewell TO, Irfan N, Gani MR, Bracken B (2013) Modest changes in fluvial style with varying accommodation in regressive alluvial-to-coastal-plain wedge: upper Cretaceous Blackhawk Formation, Wasatch Plateau, central Utah, USA. J Sediment Res 83:145–169

Hardage R (2010) Compartments can challenge logic. American Association of Petroleum Geologists, AAPG-Explorer, August 2010, pp 34–35

Hardage BA, Levey RA, Pendleton V, Simmons J, Edson R (1996) 3-D Seismic imaging and interpretation of fluvially deposited thin-bed reservoirs. In: Weimer P, Davis TL (eds) Applications of 3-D seismic data to exploration and production, vol. 42. AAPG Studies in Geology, pp 27–34

He W, Anderson RN, Xu L, Boulanger A, Meadow B, Neal R (1996) 4D seismic monitoring grows as production tool. Oil Gas J 20:41–46

Heinz J, Aigner T (2003) Three-dimensional GPR analysis of various Quaternary gravel-bed braided river deposits (southwestern Germany). In: Bristow CS, Jol HM (eds) Ground penetrating radar in sediments, vol. 211. Geological Society of London, Special Publication, pp 99–110

Heinz J, Kleineidam S, Teutsch G, Aigner T (2003) Heterogeneity patterns of Quaternary glaciofluvial gravel bodies (SW-Germany): application to hydrogeology. Sediment Geol 158:1–24

Heller PL, Angevine CL, Winslow NS, Paola C (1988) Two-phase stratigraphic model of foreland-basin sequences. Geology 16:501–504

Hentz TF, Zeng H (2003) High-frequency Miocene sequence stratigraphy, offshore Louisiana: cycle framework and influence on production distribution in a mature shelf province. Am Assoc Petr Geol B 87:197–230

Heritier FE, Lossel P, Wathne E (1980) Frigg field—large submarine fan trap in lower Eocene rocks of the Viking Graben, North Sea. In: Halbouty MT (ed) Giant oil and gas fields of the decade 1968–1978, vol. 30. American Association of Petroleum Geologists Memoir, pp 59–80

Herrero A, Alonso-Gavilán G, Colenero JR (2010) Depositional sequences in a foreland basin (north-western domain of the continental Duero basin, Spain). Sediment Geol 223:235–264

Hesselink AW, Weerts HJT, Berendsen HJA (2003) Alluvial architecture of the human-influenced River Rhine, The Netherlands. Sediment Geol 161:229–248

Hickin EJ (1983) River channel changes: retrospect and prospect. In: Collinson JD, Lewin J (eds) Modern and ancient fluvial systems, vol. 6. International Association of Sedimentologists, Special Publication, pp 61–83

Hillel D (1991) Lash of the dragon. Natural History, August 1991, pp 29–37

Hillier RD, Marriott SB, Williams BPJ, Wright VP (2007) Possible climate variability in the Lower Old Red Sandstone Conigar Pit Sandstone member (early Devonian), South Wales, UK. Sediment Geol 202:35–57

Hirst JPP (1991) Variations in alluvial architecture across the Oligo-Miocene Huesca fluvial system, Ebro basin, Spain. In: Miall AD, Tyler N (eds) The three-dimensional facies architecture of terrigenous clastic sediments, and its implications for hydrocarbon discovery and recovery, vol. 3. Society of Economic Paleontologists and Mineralogists Concepts in Sedimentology and Paleontology, pp 111–121

Hofmann MH, Wroblewski A, Boyd R (2011) Mechanisms controlling the clustering of fluvial channels and the compensational stacking of cluster belts. J Sediment Res 81:670–685

Holbrook JM (2001) Origin, genetic interrelationships, and stratigraphy over the continuum of fluvial channel-form bounding surfaces: an illustration from middle Cretaceous strata, southeastern Colorado. Sediment Geol 144:179–222

Holbrook J (2010) Valleys that never were: time surfaces versus stratigraphic surfaces—Discussion. J Sediment Res 80:2–3

Holbrook JM, Bhattacharya JP (2012) Reappraisal of the sequence boundary in time and space: case and considerations for an SU (subaerial unconformity) that is not a sediment bypass surface, a time barrier, or an unconformity. Earth Sci Rev 113:271–302

Holbrook J, Scott RW, Oboh-Ikuenobe FE (2006) Base-level buffers and buttresses: a model for upstream versus downstream control on fluvial geometry and architecture within sequences. J Sediment Res 76:162–174

Hornung J, Aigner T (1999) Reservoir and aquifer characterization of fluvial architectural elements: Stubens and stein, Upper Triassic, southwest Germany. Sediment Geol 129:215–280

Horton BK, Constenius KN, DeCelles PG (2004) Tectonic control on coarse-grained foreland-basin sequences: an example from the Cordilleran foreland basin, Utah. Geology 32:637–640

Hovius N (1998) Controls on sediment supply by larger rivers. In: Shanley KW, McCabe PJ (eds) Relative role of eustasy, climate, and tectonism in continental rocks, vol. 59. Society for Sedimentary Geology (SEPM), Special Publication, pp 3–16

Hubbard SM, Smith DG, Nielsen H, Leckie DA, Fustic M, Spencer RJ, Bloom L (2011) Seismic geomorphology and sedimentology of a tidally influenced river deposit, Lower Cretaceous Athabasca oil sands, Alberta, Canada. Am Assoc Petr Geol B 95:1123–1145

Huggenberger P (1993) Radar facies: recognition of facies patterns and heterogeneities within Pleistocene Rhine gravels, NE Switzerland. In: Best JL, Bristow CS (eds) Braided rivers, vol. 75. Geological Society, London, Special Publication, pp 163–176

Hutchison CS (1989) Geological evolution of south-east Asia, vol. 13. Clarendon Press, Oxford, Oxford Monographs on geology and geophysics, 368 p

Jackson RG II (1975) Hierarchical attributes and a unifying model of bed forms composed of cohesionless material and produced by shearing flow. Geol Soc Am Bull 86:1523–1533

Jerolmack DJ, Mohrig D (2007) Conditions for branching in depositional rivers. Geology 35:463–466

Jerolmack DJ, Paola C (2010) Shredding of environmental signals by sediment transport. Geophys Res Lett 37(L10401):5 p

Yangquan Jiao, Jiaxin Yan, Sitian Li, Ruiqi Yang, Fengjiang Lang, Shengke Yang (2005) Architectural units and heterogeneity of channel reservoirs in Karamay Formation, outcrop area of Karamay oil field, Junggar basin, northwest China. Am Assoc Petr Geol B 89:529–545

Jol HM, Bristow CS (2003) GPR in sediments: advice on data collection, basic processing and interpretation, a good practice guide. In: Bristow CS (ed) Ground penetrating radar in sediments, vol. 211. Geological Society, London, Special Publication, pp 9–27

Jones HL, Hajek EA (2007) Characterizing avulsion stratigraphy in ancient alluvial deposits. Sediment Geol 202:124–137

Jones LS, Schumm SA (1999) Causes of avulsion: an overview. In: Smith ND, Rogers J (eds) Fluvial sedimentology VI, vol. 28. International Association of Sedimentologists, Special Publication, pp 171–178

Kallmeier E, Breitkreuz C, Kiernowski H, Geißler M (2010) Issues associated with the distinction between climatic and tectonic controls on Permian alluvial fan deposits from the Kotzen and Barnim basins (North German Basin). Sediment Geol 223:15–34

Kamola D, Huntoon JE (1995) Repetitive stratal patterns in a foreland basin sandstone and their possible tectonic significance. Geology 23:177–180

Karssenberg D, Törnqvist TE, Bridge JS (2001) Conditioning a process-based model of sedimentary architecture to well data. J Sediment Res 71:868–879

Kaufman RL, Kabir CS, Abdul-Rahman B, Quttainah R, Dashti H, Pederson JM, Moon MS (2000) Characterizing the Greater Burgan field with geochemical and other field data, vol. 3. SPE Reservoir Evaluation and Engineering, pp 118–126 (SPE paper 62516)

Kesel RH, Dunne KC, McDonald RC, Allison KR, Spicer BE (1974) Lateral erosion and overbank deposition on the Mississippi River in Louisiana caused by 1973 flooding. Geology 2:461–464

Kim W, Paola C (2007) Long-period cyclic sedimentation with constant tectonic forcing in an experimental relay ramp. Geology 35:331–334

King PB (1959) The evolution of North America. Princeton University Press, Princeton, 190 p

Klingbeil R, Kleineidam S, Asprion U, Aigner T, Teutsch G (1999) Relating lithofacies to hydrofacies: outcrop-based hydrogeological characterisation of Quaternary gravel deposits. Sediment Geol 129:299–310

Knighton D (1998) Fluvial forms and processes: a new perspective. Arnold, London, 383 p

Knighton AD, Nanson GC (2000) Waterhole form and process in the anastomosing channel system of Cooper Creek, Australia. Geomorphology 35:101–117

Knox JC (2007) The Mississippi river system. In: Gupta A (ed) Large rivers: geomorphology and management. John Wiley and Sons, Chichester, pp 145–182

Komatsubara J (2004) Fluvial architecture and sequence stratigraphy of the Eocene to Oligocene Iwaki Formation, northeast Japan: channel-fills related to the sea-level change. Sediment Geol 168:109–123

Kraus MJ (1987) Integration of channel and floodplain suites, II. Vertical relations of alluvial paleosols. J Sediment Petrol 57:602–612

Kraus MJ, Aslan A (1999) Palaeosol sequences in floodplain environments: a hierarchical approach. In: Thiry M, Simon-Coincon R (eds) Palaeoweathering, palaeosurfaces and related continental deposits, vol. 27. International Association of Sedimentologists, Special Publication, pp 303–321

Kraus MJ, Wells TM (1999). Recognizing avulsion deposits in the ancient stratigraphical record. In: Smith ND, Rogers J (eds) Fluvial sedimentology VI, vol. 28, Special Publication of the International Association of Sedimentologists, pp 251–268

Krystinik LF, DeJarnett BB (1995) Lateral variability of sequence stratigraphic framework in the Campanian and Lower Maastrichtian of the Western Interior Seaway. In: Van Wagoner JC, Bertram GT (eds) Sequence stratigraphy of foreland basins, vol. 64. American Association of Petroleum Geologists Memoir, pp 11–25

Labourdette R (2011) Stratigraphy and static connectivity of braided fluvial deposits of the lower Escanilla Formation, south central Pyrenees, Spain. Am Assoc Petr Geol 95:585–617

Lane EW (1955) The importance of fluvial morphology in hydraulic engineering. Am Soc Civil Eng Proc 81(745):1–17

Larue DK, Friedmann F (2005) The controversy concerning stratigraphic architecture of channelized reservoirs and recovery by waterflooding. Petrol Geosci 11:131–146

Larue DK, Hovadik J (2006) Connectivity of channelized reservoirs: a modelling approach. Petrol Geosci 12:291–308

Larue DK, Hovadik J (2008) Why is reservoir architecture an insignificant uncertainty in many appraisal and development studies of clastic channelized reservoirs? J Petrol Geol 31:337–366

Laurin J, Sageman BB (2007) Cenomanian-Turonian coastal record in SW Utah, USA: orbital-scale transgressive-regressive events during oceanic anoxic event II. J Sediment Res 77:731–756

Leckie DA (1994) Canterbury Plains, New Zealand—implications for sequence stratigraphic models. Am Assoc Petr Geol B 78:1240–1256

Leclair SF (2011) Interpreting fluvial hydromorphology from the rock record: large-river peak flow leaves no clear signature. In: Davidson SK, Leleu S, North CP (eds) From river to rock record, vol. 97. Society for Sedimentary Geology (SEPM), Special Publication, pp. 113–124

Leclair SF, Bridge JS (2001) Quantitative interpretation of sedimentary structures formed by river dunes. J Sediment Res 71:713–716

Leeder MR (1973) Fluviatile fining-upward cycles and the magnitude of palaeochannels. Geol Mag 110:265–276

Leeder MR (1978) A quantitative stratigraphic model for alluvium, with special reference to channel deposit density and interconnectedness. In: Miall AD (ed) Fluvial sedimentology, vol. 5. Canadian Society of Petroleum Geologists Memoir, pp 587–596

Leeder MR (1993) Tectonic controls upon drainage basin development, river channel migration and alluvial architecture: implications for hydrocarbon reservoir development and characterization. In: North CP, Prosser DJ (eds) Characterization of fluvial and aeolian reservoirs, vol. 73. Geological Society, London, Special Publication, pp 7–22.

Leeder MR, Alexander J (1987) The origin and tectonic significance of asymmetric meander belts. Sedimentology 34:217–226

Leeder MR, Gawthorpe RL (1987) Sedimentary models for extensional tilt-block/half-graben basins. In: Coward MP, Dewey JF, Hancock PL (eds) Continental extension tectonics, vol. 28. Geological Society, London, Special Publication, pp 139–152

Leeder MR, Stewart MD (1996) Fluvial incision and sequence stratigraphy: alluvial responses to relative sea-level fall and their detection in the geological record. In: Hesselbo SP, Parkinson DN (eds) Sequence stratigraphy in British Geology, vol. 103. Geological Society, London, Special Publication, pp 25–39

Leier AL, DeCelles PG, Pelletier JD (2005) Mountains, monsoons and megafans. Geology 33:289–292

Leopold LB, Langbein WB (1966). River meanders. Scientific American, New York, p 214

Leopold LB, Wolman MG (1957) River channel patterns; braided, meandering, and straight. US Geological Survey Professional Paper 282-B.

Leopold LB, Wolman MG, Miller JP (1964) Fluvial processes in geomorphology. W.H. Freeman and Co., San Francisco, 522 p

Lesmes DP, Decker SM, Roy DC (2002) Amultiscale radar-stratigraphic analysis of fluvial aquifer heterogeneity. Geophysics 67:1452–1464

Li S, Finlayson B (1993) Flood management on the lower Yellow River: hydrological and geomorphological perspectives. Sediment Geol 85:285–296

Liu SF, Nummedal D, Yin PG, Luo HJ (2005) Linkage of Sevier thrusting episodes and Late Cretaceous foreland basin megasequences across southern Wyoming (USA). Basin Res 17:487–506

Long DGF (2006) Architecture of pre-vegetation sandy-braided perennial and ephemeral river deposits in the Paleoproterozoic Athabasca Group, northern Saskatchewan, Canada as indicators of Precambrian fluvial style. Sediment Geol 190:71–95

Long DGF (2011) Architecture and depositional style of fluvial systems before land plants: a comparison of Precambrian, early paleozoic and modern river deposits. In: Davidson SK, Leleu S, North CP (eds) From river to rock record, vol. 97. Society for Sedimentary Geology (SEPM), Special Publication, pp 37–61

López-Gómez J, Arche A, Vargas H, Marzo M (2010) Fluvial architecture as a response to two-layer lithospheric subsidence during the Permian and Triassic in the Iberian basin, eastern Spain. Sediment Geol 223:320–333

Lucas SG (1997) The upper Triassic Chinle Group: western United States: a nonmarine standard for Late Triassic time. In: Morales M (ed) Aspects of mesozoic geology and paleontology of the Colorado Plateau, vol. 59. Museum of Northern Arizona Bulletin, pp 27–50

Lucas SG, Heckert AB, Estep JW, Anderson O (1997) Stratigraphy of the upper triassic Chinle group, four corners region. In: Anderson OJ, Kues B, Lucas SG (eds) Mesozoic geology and paleontology of the four corners region. New Mexico Geological Society, Guidebook, pp 81–108

Lukie TD, Ardies GW, Dalrymple RW, Zaitlin BA (2002) Alluvial architecture of the Horsefly unit (Basal Quartz) in southern Alberta and northern Montana: influence of accommodation changes and contemporaneous faulting. B Can Petrol Geol 50:73–91

Lunt IA, Bridge JS (2004) Evolution and deposits of a gravelly braid bar, Sagavanirktok River, Alaska. Sedimentology 51:415–432

Lunt IA, Bridge JS, Tye RS (2004) A quantitative, three-dimensional depositional model of gravelly braided rivers. Sedimentology 51:377–414

Mack GH, Leeder M, Perez-Arlucea M, Durr M (2011) Tectonic and climatic conrols on Holocene channel migration, incision and terrace formation by the Rio Grande in the Palomas half graben, southern Rio Grande rift, USA. Sedimentology 58:1065–1086

Mack GH, Madoff RD (2005) A test of models of fluvial architecture and palaeosol development: Camp Rice Formation (Upper Pliocene-Lower Pleistocene), southern Rio Grande Rift, New Mexico, USA. Sedimentology 52:191–211

Mack GH, Seager WR, Leeder MR, Perez-Arlucea M, Salyards SL (2006) Pliocene and Quaternary history of the Rio Grande, the axial river of the southern Rio Grande rift, New Mexico, USA. Earth Sci Rev 79:141–162

Mackey SD, Bridge JS (1995) Three-dimensional model of alluvial stratigraphy: theory and application. J Sediment Res B65:7–31

Makaske B (2001) Anastomosing rivers: a review of their classification, origin and sedimentary products. Earth Sci Rev 53:149–196

Marriott SB (1999) The use of models in the interpretation of the effects of base-level change on alluvial architecture. In: Smith ND, Rogers J (eds) Fluvial sedimentology VI, vol. 28. International Association of Sedimentologists, Special Publication, pp 271–281

Marriott SB, Wright VP, Williams BPJ (2005) A new evaluation of fining-upward sequences in a mud-rock dominated succession of the Lower Old Red Sandstone of South Wales, UK. In: Blum MD, Marriott SB, Leclair SF (eds) Fluvial sedimentology VII, vol. 35. International Association of Sedimentologists, Special Publication, pp 517–529

Martin JH (1993) A review of braided fluvial hydrocarbon reservoirs: the petroleum engineer's perspective. In: Best JL, Bristow CS (eds) Braided rivers, vol. 75. Geological Society, London, Special Publication, pp 333–367

Martinius AW (1996) The sedimentological characterization of labyrinthine fluvial reservoir analogues. Doctoral thesis, Delft University of Technology, 300 p

Martinius AW (2000) Labyrinthine facies architecture of the Tórtola fluvial system and controls on deposition (late Oligocene-early Miocene, Loranca Basin, Spain). J Sediment Res 70:850–867

Martinsen OJ, Ryseth A, Helland-Hansen W, Flesche H, Torkildsen G, Idel S (1999) Stratigraphic base level and fluvial architecture: Ericson Sandstone (Campanian), Rock springs uplift, SW Wyoming, USA. Sedimentology 46:235–259

Matthews MD, Perlmutter MA (1994) Global cyclostratigraphy: an application to the Eocene Green River Basin. In: de Boer PL, Smith DG (eds) Orbital forcing and cyclic sequences, vol. 19. International Association of Sedimentologists, Special Publication, pp 459–481
Maynard JR, Feldman HR, Always R (2010) From bars to valleys: the sedimentology and seismic geomorphology of fluvial to estuarine incised-valley fills of the Grand Rapids Formation (Lower Cretaceous), Iron River field, Alberta, Canada. J Sediment Res 80:611–638
McCarthy PJ (2002) Micromorphology and development of interfluve paleosols: a case study from the Cenomanian Dunvegan formation, NE British Columbia, Canada. B Can Petrol Geol 50:158-177
McCarthy PJ, Plint AG (1998) Recognition of interfluve sequence boundaries: integrating paleopedology and sequence stratigraphy. Geology 26:387–390
McCarthy PJ, Plint AG (2003) Spatial variability of palaeosols across Cretaceous interfluves in the Dunvegan Formation, NE British Columbia, Canada: palaeohydrological, palaeogeomorphological and stratigraphic implications. Sedimentology 50:1187–1220
McCarthy TS, Ellery WN, Stanistreet IG (1992) Avulsion mechanism on the Okavango fan, Botswana: the control of a fluvial system by vegetation. Sedimentology 39:779–795
McLaurin BT, Steel RJ (2007) Architecture and origin of an amalgamated fluvialsheetsand, lower Castlegate formation, BookCliffs, Utah. Sediment Geol 197:291–311
McMechan GA, Gaynor GC, Szerbiak RB (1997) Use of ground-penetrating radar for 3-D sedimentological characterization of clastic reservoir analogs. Geophysics 62:786–796
McMillan NJ (1973) Shelves of Labrador Sea and Baffin Bay, Canada. In: The future petroleum provinces of Canada, their geology and potential. Can Soc Petrol Memoir 1:473–517
Medwedeff DA (1989) Growth fault-bend folding at Southeast Lost Hills, San Joaquin valley, California. Am Assoc Petr Geol B 73:54–67
Mezghani M, Fornel A, Langlais V, Lucet N (2004) History matching and quantitative use of 4D seismic data for an improved reservoir characterization. Society of Petroleum Engineers, paper SPE 90420.
Miall AD (1980) Cyclicity and the facies model concept in geology. B Can Petrol Geol 28:59–80
Miall AD (1981) Alluvial sedimentary basins: tectonic setting and basin architecture. In: Miall AD (ed) Sedimentation and tectonics in alluvial basins, vol. 23. Geological Association of Canada Special Paper, pp 1–33
Miall AD (1985) Architectural-element analysis: a new method of facies analysis applied to fluvial deposits. Earth Sci Rev 22:261–308
Miall AD (1988) Reservoir heterogeneities in fluvial sandstones: lessons from outcrop studies. Am Assoc Petr Geol B 72:682–697
Miall AD (1991a) Hierarchies of architectural units in clastic rocks, and their relationship to sedimentation rate. In: Miall AD, Tyler N (eds) The three-dimensional facies architecture of terrigenous clastic sediments, and its implications for hydrocarbon discovery and recovery, vol. 3 Society of Economic Paleontologists and Mineralogists, Concepts in Sedimentology and Paleontology, pp 6–12
Miall AD (1991b) Stratigraphic sequences and their chronostratigraphic correlation. J Sediment Petrol 61:497–505
Miall AD (1994) Reconstructing fluvial macroform architecture from two-dimensional outcrops: examples from the Castlegate Sandstone, Book Cliffs, Utah. J Sediment Res B64:146–158
Miall AD (1996) The geology of fluvial deposits: sedimentary facies, basin analysis and petroleum geology. Springer-Verlag, Inc., Heidelberg, 582 p
Miall AD (2002) Architecture and sequence stratigraphy of Pleistocene fluvial systems in the Malay Basin, based on seismic time-slice analysis. Am Assoc Petr Geol B 86:1201–1216
Miall AD (2006a) Reconstructing the architecture and sequence stratigraphy of the preserved fluvial record as a tool for reservoir development: a reality check. Am Assoc Petr Geol B 90:989–1002
Miall AD (2006b) How do we identify big rivers, and how big is big? Sediment Geol 186:39–50
Miall AD (2010a) Alluvial deposits. In: James NP, Dalrymple RW (eds) Facies models 4, Geological Association of Canada, St. John's, Newfoundland, GEO text 6, pp 105–137 (50)
Miall AD (2010b) The geology of stratigraphic sequences. 2nd edn. Springer-Verlag, Berlin, 522 p

Miall AD, in press. A new uniformitarianism: stratigraphy as just a set of "frozen accidents". In: Smith DG, Bailey RJ, Burgess P, Fraser A (eds) Strata and time. Geological Society, London, Special Publication

Miall AD, Arush M (2001a) The Castlegate Sandstone of the Book Cliffs, Utah: sequence stratigraphy, paleogeography, and tectonic controls. J Sediment Res 71:536–547

Miall AD, Arush M (2001b) Cryptic sequence boundaries in braided fluvial successions. Sedimentology 48(5):971–985

Miall AD, Jones B (2003) Fluvial architecture of the Hawkesbury Sandstone (Triassic), near Sydney, Australia. J Sediment Res 73:531–545

Middleton GV (ed) (1965) Primary sedimentary structures and their hydrodynamic interpretation. Society of Economic Paleontologists and Mineralogists Special Publication, vol. 12, 265 p

Miller KG, Kominz MA, Browning JV, Wright JD, Mountain GS, Katz ME, Sugarman PJ, Cramer BS, Christie-Blick N, Pekar SF (2005a) The Phanerozoic record of global sea-level change. Science 310:1293–1298

Miller KG, Wright JD, Browning JV (2005) Visions of ice sheets in a greenhouse world. Marine Geol 217:215–231

Mohrig D, Heller PL, Paola C, Lyons WJ (2000) Interpreting avulsion process from ancient alluvial sequences: Guadalope–Matarranya system (northern Spain) and Wasatch Formation, (western Colorado). Geol Soc Am Bull 112(12):1787–1803

Molenaar CM, Rice DD (1988) Cretaceous rocks of the Western Interior Basin. In: Sloss LL (ed) Sedimentary cover—North American Craton: US, The Geology of North America, vol. D-2. Geological Society of America, Boulder, CO, pp 77–82

Morend D, Pugin A, Gorin GE (2002) High-resolution imaging of outcrop-scale channels and an incised valley system within the fluvial-dominated Lower Freshwater Molasse (Aquitanian, western Swiss Molasse Basin). Sediment Geol 149:245–264

Morgan JP, McKintire WG (1959) Quaternary geology of the Bengal Basin, East Pakistan and India. Geol Soc Am Bull 70:319–342

Morozova GS, Smith ND (1999) Holocene avulsion history of the lower Saskatchewan fluvial system, Cumberland Marshes, Saskatchewan-Manitoba, Canada. In: Smith ND, Rogers J (eds) Fluvial sedimentology VI, vol. 28. International Association of Sedimentologists, Special Publication, pp 231–249

Morozova GS, Smith ND (2000) Holocene avulsion styles and sedimentation patterns of the Saskatchewan river, Cumberland Marshes, Canada. Sediment Geol 130:81–105

Mossop GD, Flach PD (1983) Deep channel sedimentation in the lower Cretaceous McMurray formation, Athabasca oil sands, Alberta. Sedimentology 30:493–509

Nanson GC (1980) Point bar and floodplain formation of the meandering Beatton River, northeastern British Columbia, Canada. Sedimentology 27:3–30

Nanson GC, Croke JC (1992) A genetic classification of floodplains. Geomorphology 4:459–486

Nanson GC, Knighton AD (1996) Anabranching rivers: their cause, character and classification. Earth Surf Proc Land 21:217–239

Nanz RH Jr (1954) Genesis of Oligocene sandstone reservoir, Seeligson field, Jim Wells and Kleberg Counties, Texas. Am Assoc Petr Geol B 38:96–117

Neal A (2004) Ground-penetrating radar and its uses in sedimentology: principles, problems and progress. Earth Sci Rev 66:261–330

Nelson BW (1970) Hydrography, sediment dispersal and recent historical development of the Po River delta, Italy. In: Morgan JP, Shaver RH (eds) Deltaic sedimentation, modern and ancient, vol. 15. Society of Economic Paleontologists and Mineralogists Special Publication, pp 152–184

Nijman W (1998) Cyclicity and basin axis shift in a piggyback basin: toward modeling of the Eocene Tremp-Ager basin, south Pyrenees, Spain. In: Mascle A, Puigdefàbregas C, Luterbacher HP, Fernàndez M (eds) Cenozoic foreland basins of western Europe, vol. 134. Geological Society, London, Special Publication, pp 135–162

North CP (1996) The prediction and modelling of subsurface fluvial stratigraphy. In: Carling PA, Dawson MR (eds) Advances in fluvial dynamics and stratigraphy. John Wiley and Sons, Chichester , pp 395–508

North CP, Davidson SK (2012) Unconfined alluvial flow processes: recognition and interpretation of their deposits and the significance for palaeogeographic reconstruction. Earth Sci Rev 111:199–223

North CP, Nanson GC, Fagan SD (2007) Recognition of the sedimentary architecture of dry-landanabranching (anastomosed) rivers. J Sediment Res 77:925–938

North CP, Taylor KS (1996) Ephemeral-fluvial deposits: integrated outcrop and simulation studies reveal complexity. Am Assoc Petr Geol B 80:811–830

O'Leary J, Hill GS (1989) Tertiary basin development in the southern Central Plains, Thailand. In: International Symposium on Intermontane Basins, Geology and Resources, Chiang Mai, Thailand, pp 254–264

Olsen PE (1990) Tectonic, climatic, and biotic modulation of lacustrine ecosystems—examples from Newark Supergroup of eastern North America. In: Katz BJ (ed) Lacustrine basin exploration: case studies and modern analogs, vol. 50. American Association of Petroleum Geologists Memoir, pp 209–224

Olsen T, Steel RJ, Høgseth K, Skar T, Røe S-L (1995) Sequential architecture in a fluvial succession: sequence stratigraphy in the upper Cretaceous Mesaverde Group, Price Canyon, Utah. J Sediment Res B65:265–280

Oreskes N, Shrader-Frechette K, Belitz K (1994) Verification, validation, and confirmation of numerical models in the earth sciences. Science 263:641–646

Ori GG (1979) Barre di meandrenellealluvionighiaiose del fiume Reno (Bologna). Bull Soc Geol Ital 98:35–54

Paola C (2000) Quantitative models of sedimentary basin filling. Sedimentology 47:121–178

Paola C, Straub K, Mohrig D, Reinhardt L (2009) The "unreasonable effectiveness" of stratigraphic and geomorphic experiments. Earth Sci Rev 97:1–43

Parker G (1976) On the cause and characteristic scales of meandering and braiding in rivers. J Fluid Mech 76:457–480

Peper T, Beekman F, Cloetingh S (1992) Consequences of thrusting and intraplate stress fluctuations for vertical motions in foreland basins and peripheral areas. Geophys J Int 111:104–126

Perlmutter MA, Matthews MD (1990) Global cyclostratigraphy—a model. In: Cross TA (ed) Quantitative dynamic stratigraphy. Prentice Hall, Englewood Cliffs, pp 233–260

Petter AL (2011) Reconstructing backwater reaches of paleorivers and their influence on fluvial facies distribution, Campanian Lower Castlegate Sandstone, Utah. Geological Society of America Annual Meeting, Minneapolis, Paper 149-12

Pettijohn FJ (1957) Sedimentary rocks. Harper & Bros., New York, 718 p

Pietras JT, Carroll AR (2006) High-resolution stratigraphy of an underfilled lake basin: Wilkins Peak Member, Eocene Green River Formation, Wyoming, USA. J Sediment Res 76:1197–1214

Platt NH, Keller B (1992) Distal alluvial deposits in a foreland basin setting—the Lower Freshwater Molasse (Lower Miocene), Switzerland: sedimentology, architecture and palaeosols. Sedimentology 39:545–565

Plint AG (1991) High-frequency relative sea-level oscillations in Upper Cretaceous shelf clastics of the Alberta foreland basin: possible evidence for a glacio-eustatic control? In: Macdonald DIM (ed) Sedimentation, tectonics and eustasy: sea-level changes at active margins, vol. 12. International Association of Sedimentologists, Special Publication, pp 409–428

Plint AG (2002) Paleovalley systems in the upper Cretaceous Dunvegan Formation, Alberta and British Columbia. B Can Petrol Geol 50:277–298

Plint AG, Kreitner MA (2007) Extensive thin sequences spanning Cretaceous foredeep suggest high-frequency eustatic control: late Cenomanian Western Canada foreland basin. Geology 35:735–738

Plint AG, McCarthy PJ, Faccini UF (2001) Nonmarine sequence stratigraphy: updip expression of sequence boundaries and systems tracts in a high-resolution framework: Cenomanian Dunvegan formation, Alberta foreland basin, Canada. Am Assoc Petr Geol B 85:1967–2001

Plint AG, Wadsworth JA (2003) Sedimentology and palaeogeomorphology of four large valley systems incising delta plains, western Canada Foreland Basin: implications for mid-Cretaceous sea-level changes. Sedimentology 50:1147–1186

Posamentier HW (2001) Lowstand alluvial bypass systems: incised vs. unincised. Am Assoc Petr Geol B 85:1771–1793

Posamentier HW, Allan GP, James DP (1992) High-resolution sequence stratigraphy—the East Coulee Delta, Alberta. J Sediment Petrol 62:310–317

Posamentier HW, Davies RJ, Cartwright JA, Wood L (2007) Seismic geomorphology—an overview. In: Davies RJ, Posamentier HW, Wood LJ, Cartwright JA (eds) Seismic geomorphology: applications to hydrocarbon exploration and production, vol. 277. Geological Society, London, Special Publication, pp 1–14

Posamentier HW, Jervey MT, Vail PR (1988) Eustatic controls on clastic deposition I–conceptual framework. In: Wilgus CK, Hastings BS, Kendall CGStC, Posamentier HW, Ross CA, Van Wagoner JC (eds) Sea-level research: an integrated approach, vol. 42. Society of Economic Paleontologists and Mineralogists, Special Publication, pp 109–124

Posamentier HW, Vail PR (1988) Eustatic controls on clastic deposition II–sequence and systems tracts models. In: Wilgus CK, Hastings BS, Kendall CGStC, Posamentier HW, Ross CA, Van Wagoner JC (eds) Sea-level research: an integrated approach, vol. 42. Society of Economic Paleontologists and Mineralogists Special Publication, pp 125–154

Potter PE (1967) Sand bodies and sedimentary environments: a review. Am Assoc Petr Geol B 51:337–365

Potter PE (1978) Significance and origin of big rivers. J Geol 86:13–33

Pranter MJ, Cole RD, Panjaitan H, Sommer NK (2009) Sandstone-body dimension in a lower coastal-plain setting: lower Williams Fork formation, Coal Canyon, Piceance Basin, Colorado. Am Assoc Petr Geol 93:1379–1401

Prochnow SJ, Atchley SC, Boucher TE, Nordt LC, Hudec MR (2006) The influence of salt withdrawal subsidence on palaeosol maturity and cyclic fluvial deposition in the Upper Cretaceous Chinle formation, Castel Valley, Utah. Sedimentology 53:1319–1345

Puigdefábregas C (1973) Miocene point-bar deposits in the Ebro Basin, northern Spain. Sedimentology 20:133–144

Püspöki Z, Demeter G, Tóth-Makk Á, Kozák M, Dávid Á, Virág M, Kovács-Pálffy P, Kónya P, Gyuricza Gy, Kiss J, McIntosh RW, Forgács Z, Buday T, Kovács Z, Gombos T, Kiummer I (2013) Tectonically-controlled Quaternary intracontinental fluvial sequence development in the Nyírség-Pannonian Basin, Hungary. Sediment Geol 283:34–56

Rainbird RH (1992) Anatomy of a large-scale braid-plain quartzarenite from the Neoproterozoic Shaler Group, Victoria Island, Northwest Territories, Canada. Canad J Earth Sci 29:2537–2550

Rainbird RH, McNicoll VJ, Heaman LM, Abbott JG, Long DGF, Thorkelson DJ (1997) Pan-continental river system draining Grenville Orogen recorded by U-Pb and Sm-Nd geochronology of Neoproterozoic quartzarenites and mudrocks, northwestern Canada. J Geol 105:1–17

Rannie WF (1990) The Portage La Prairie Floodplain Fan. In: Rachocki AH, Church M (eds) Alluvial fans: a field approach.John Wiley and Sons Ltd, Chichester, pp 179–193

Ramos A, Sopeña A (1983) Gravel bars in low-sinuosity streams (Permian and Triassic, central Spain). In: Collinson JD, Lewin J (eds) Modern and ancient fluvial systems, vol. 6. International Association of Sedimentologists, Special Publication, pp 301–312

Ramos A, Sopeña A, Perez-Arlucea M (1986) Evolution of Buntsandstein fluvial sedimentation in the northwest Iberian Ranges (Central Spain). J Sediment Petrol 56:862–875

Refunjol BT, Lake LW (1999) Reservoir characterization based on tracer response and rank analysis of production and injection rates. In: Schatzinger R, Jordan J (eds) Reservoir characterization: recent advances, vol. 71. American Association of Petroleum Geologists, Memoir, pp 209–218

Reijenstein HM, Posamentier HW, Bhattacharya JP (2011) Seismic geomorphology and high-resolution seismic stratigraphy of inner-shelf fluvial, estuarine, deltaic, and marine sequences, Gulf of Thailand. Am Assoc Petr Geol B 95:1959–1990

Retallack GJ Soils of the past: an introduction to paleopedology. 2nd edn. Blackwell Science, Oxford, 404 p

Riba O (1976) Syntectonic unconformities of the Alto Cardener, Spanish Pyrenees, a genetic interpretation. Sediment Geol 15:213–233

Richards K, Chandra S, Friend PF (1993) Avulsive channel systems: characteristics and examples. In: Best JL, Bristow CS (eds) Braided rivers, vol. 75. Geological Society, London, Special Publication, pp 195–203

Ricketts BD (2008) Cordilleran sedimentary basins of western Canada record 180 million years of terrane accretion. In: Miall AD (ed) The sedimentary basins of the United States and Canada, vol. 5. Sedimentary basins of the World, K.J. Hsü, Series Editor, Elsevier Science, Amsterdam, pp 363–394

Robinson JW, McCabe PJ (1997) Sandstone-body and shale-body dimensions in a braided fluvial system: salt wash sandstone member (Morrison Formation), Garfield County, Utah. Am Assoc Petr Geol B 81:11267–1291

Robinson RAJ, Slingerland RL (1998) Grain-size trends, basin subsidence and sediment supply in the Campanian Castlegate Sandstone and equivalent conglomerates of central Utah. Basin Res 10:109–127

Rust BR, Jones BG (1987) The Hawkesbury Sandstone south of Sydney, Australia: Triassic analogue for the deposit of a large braided river. J Sediment Petrol 57:222–233

Rust B R, Legun AS (1983) Modern anastomosing-fluvial deposits in arid central Australia, and a Carboniferous analogue in New Brunswick, Canada. In: Collinson JD, Lewin J (eds) Modern and ancient fluvial systems, vol. 6. International Association of Sedimentologists, Special Publication, pp 385–392

Ryer TA (1984) Transgressive-regressive cycles and the occurrence of coal in some Upper Cretaceous strata of Utah, USA. In: Rahmani RA, Flores RM (eds) Sedimentology of coal and coal-bearing sequences, vol. 7. International Association of Sedimentologists, Special Publication, pp 217–227

Ryseth A, Fjellbirkeland H, Osmundsen IK, Skålnes Å, Zachariassen E (1998) High-resolution stratigraphy and seismic attribute mapping of a fluvial reservoir: middle Jurassic Ness formation, Oseberg field. Am Assoc Petr Geol B 82:1627–51

Sadler PM (1981) Sedimentation rates and the completeness of stratigraphic sections. J Geol 89:569–584

Sageman BB, Rich J, Arthur MA, Birchfield GE, Dean WE (1997) Evidence for Milankovitch periodicities in Cenomanian-Turonian lithologic and geochemical cycles, western interior, USA. J Sediment Res 67:286–302

Sageman B, Rich J, Savrada CE, Bralower T, Arthur MA, Dean WE (1998). Multiple Milankovitch cycles in the Bridge Creek Limestone (Cenomanian-Turonian), Western Interior Basin. In: Arthur MA, Dean WE (eds) Stratigraphy and paleo environments of the Cretaceous Western Interior Seaway, vol. 6. SEPM Concepts in Sedimentology and Paleontology, pp 153–171

Sambrook-Smith GH, Ashworth PJ, Best JL, Woodward J, Simpson CJ (2005) The morphology and facies of sandy braided rivers: some considerations of scale invariance. In: Blum MD, Marriott SB, Leclair SF (eds) Fluvial sedimentology VII, vol. 35. International Association of Sedimentologists, Special Publication, pp 145–158

Sambrook-Smith GH, Ashworth PJ, Best JL, Woodward J, Simpson CJ (2006) The sedimentology and alluvial architecture of the sandy braided South Saskatchewan River, Canada. Sedimentology 53:413–434

Sambrook Smith GH, Best JL, Ashworth PJ, Lane SN, Parker NO, Lunt IA, Thomas RE, Simpson CJ (2010) Can we distinguish flood frequency and magnitude in the sedimentological record of rivers. Geology 38:579–582

Sarzalejo S, Hart BS (2006) Stratigraphy and lithologic heterogeneity in the Mannville Group (southeast Saskatchewan) defined by integrating 3-D seismic and log data. B Can Petrol Geol 54:138–151

Saucier RT (1974) Quaternary geology of the lower Mississippi Valley. Arkansas Archaeological Survey Research Series, vol. 6. 26 p

Saucier RT (1994) Geomorphology and Quaternary geologic history of the lower Mississippi Valley. Waterways Experiment Station, US Army Corps of Engineers, Vicksburg, Mississippi 364 p

Saucier RT (1996) A contemporary appraisal of some key Fiskian concepts with emphasis on Holocene meander belt formation and morphology. Eng Geol 45:67–86

Schumm SA (1963) A tentative classification of alluvial river channels. US Geological Survey Circular, 477

Schumm SA (1968a) Speculations concerning paleohydrologic controls of terrestrial sedimentation. Geol Soc Am Bull 79:1573–1588

Schumm SA (1968b) River adjustment to altered hydrologic regimen — Murrumbidgee River and paleochannels, Australia. US Geological Survey Professional Paper 598, 65 p

Schumm SA (1977) The fluvial system. John Wiley and Sons, New York, 338 p

Schumm SA (1979) Geomorphic thresholds: the concept and its applications. T I Brit Geogr 4:485–515

Schumm SA (1981) Evolution and response of the fluvial system, sedimentological implications. In: Ethridge FG, Flores RM (eds) Recent and ancient nonmarine depositional environments: models for exploration, vol. 31. Society of Economic Paleontologists and Mineralogists, Special Publication, pp 19–29

Schumm SA (1985) Patterns of alluvial rivers. Annu Rev Earth Pl Sc 13:5–27

Schumm SA (1993) River response to base level change: implications for sequence stratigraphy. J Geol 101:279–294

Schumm SA (1994) The variability of large alluvial rivers. American Society of Civil Engineers, New York, 467 p

Schumm SA, Dumont JF, Holbrook JM (2000) Active tectonics and alluvial rivers. Cambridge University Press, Cambridge, 276 p

Seward AC (1959) Plant life through the ages. Hafner, New York, 603 p

Shanley KW (2004) Fluvial reservoir description for a giant, low-permeability gas field: Jonah field, Green river Basin, Wyoming, USA. In: Robinson JW, Shanley KW (eds) Jonah field: case study of a tight-gas fluvial reservoir. AAPG Studies in Geology, vol. 52, pp 159–182

Shanley K, McCabe PJ (1989) Sequence-stratigraphic relationships and facies architecture of Turonian-Campanian strata, Kaiparowits Plateau, south-central Utah. AAPG Bull 73:410–411

Shanley KW, McCabe PJ (1994) Perspectives on the sequence stratigraphy of continental strata. Am Assoc Petr Geol B 78:544–568

Shanley KW, McCabe PJ, Hettinger RD (1992) Significance of tidal influence in fluvial deposits for interpreting sequence stratigraphy. Sedimentology 39:905–930

Sheets BA, Hickson TA, Paola C (2002) Assembling the stratigraphic record: depositional patterns and time-scales in an experimental alluvial basin. Basin Res 14:287–301

Sheets BA, Paola C, Kelberer JM (2007) Creation and preservation of channel-form sand bodies in an experimental alluvial system. In: Nichols G, Williams E, Paola C (eds) Sedimentary processes, environments and basins: a tribute to Peter Friend, vol. 38. International Association of Sedimentologists, Special Publication, vol. 38, pp 555–567

Shepard FP, Wanless HR (1935) Permo-Carboniferous coal series related to southern hemisphere glaciation. Science 81:521–522

Singh H, Parkash B, Gohain K (1993) Facies analysis of the Kosimegafan deposits. Sediment Geol 85:87–113

Sinha R, Bhattacharjee PS, Sangode SJ, Gibling MR, Tandon SK, Jain M, Godfrey-Smith D (2007) Valley and interfluve sediments in the Southern Ganga plains, India: exploring facies and magnetic signatures. Sediment Geol 201:386–411

Sinha R, Gibling MR, Jain V, Tandon SK (2005) Sedimentology and avulsion patterns of the anabranching Baghmati River in the Himalayan foreland basin, India. In: Blum MD, Marriott SB, Leclair SF (eds) Fluvial sedimentology VII, vol. 35. International Association of Sedimentologists, Special Publication, pp 181–196

Slingerland R, Smith ND (1998) Necessary conditions for meandering-river avulsion. Geology 26:435–438

Sloss LL (1962) Stratigraphic models in exploration. Am Assoc Petr Geol B 46:1050–1057

Sloss LL (1963) Sequences in the cratonic interior of North America. Geol Soc Am Bull 74:93–113

Smalley PC, Hale NA (1996) Early identification of reservoir compartmentalization by combining a range of conventional and novel data types. SPE Formation Evaluation, September 1996, pp 163–169 (SPE paper 30533)

Smith DG (1983) Anastomosed fluvial deposits: modern examples from western Canada. In: Collinson JD, Lewin J (eds) Modern and ancient fluvial systems, vol. 6. International Association of Sedimentologists, Special Publication, pp 155–168

Smith DG, Smith ND (1980) Sedimentation in anastomosed river systems: examples from alluvial valleys near Banff, Alberta. J Sediment Petrol 50:157–164

Smith ND, Cross TA, Dufficy JP, Clough SR (1989) Anatomy of an avulsion. Sedimentology 36:1–23

Smith ND, Slingerland RL, Pérez-Arlucea M, Morozova GS (1998) The 1870s avulsion of the Saskatchewan River. Canad J Earth Sci 35:453–466

Soreghan GS, Montanez IP (eds) (2008) Special issue of the late Paleozoic earth system: palaeogeography, palaeoclimatology. Palaeoecology 268(3–4):310 p.

Stephen KD, Dalrymple M (2002) Reservoir simulations developed from an outcrop of incise valley fill strata. Am Assoc Petr Geol B 86:797–822

Stephens M (1994) Architectural element analysis within the Kayenta Formation (Lower Jurassic) using ground-probing radar and sedimentological profiling, southwestern Colorado. Sediment Geol 90:179–211

Stevaux JC, Souza IA (2004) Floodplain construction in an anastomosed river. Quat Int 114:55–65

Stockmal GS, Cant DJ, Bell JS (1992) Relationship of the stratigraphy of the Western Canada foreland basin to Cordilleran tectonics: insights from geodynamic models. In: Macqueen RW, Leckie DA (eds) Foreland basin and fold belts, vol. 55. American Association of Petroleum Geologists Memoir, pp 107–124

Stølum H-H (1996) River meandering as a self-organizing process. Science 271:1710–1713

Stouthamer E (2001a) Holocene avulsions in the Rhine-Meuse delta, vol. 283. Netherlands Geographical Studies, The Netherlands, 224 p

Stouthamer E (2001b) Sedimentary products of avulsions in the Holocene Rhine-Meuse delta, The Netherlands. Sediment Geol 145:73–92

Stouthamer E, Berendsen HJA (2007) Avulsion: the relative roles of autogenic and allogenic processes. Sediment Geol 198:309–325

Stouthamer E, Cohen KM, Gouw MJP (2011) Avulsion and its implication for fluvial-deltaic architecture: insights from the Holocene Rhine-Meuse delta. In: Davidson SK, Leleu S, North C P (eds) From river to rock record, vol. 97. Society for Sedimentary Geology (SEPM), Special Publication, pp 215–231

Straub KM, Paola C, Mohrig D, Wolinsky MA, George T (2009) Compensational stacking of channelized sedimentary deposits. J Sediment Res 79:673–688

Strong N, Paola C (2008) Valleys that never were: time surfaces versus stratigraphic surfaces. J Sediment Res 78:579–593

Strong N, Sheets B, Hickson T, Paola C (2005) A mass-balance framework for quantifying downstream changes in fluvial architecture. In: Blum MD, Marriott SB, Leclair SF (eds) Fluvial sedimentology VII, vol. 35. International Association of Sedimentologists, Special Publication, pp 243–253

Sun J, Li Y, Zhang Z, Fu B (2010) Magnetostratigraphic data on Neogene growth folding in the foreland basin of the southern Tianshan Mountains. Geology 37:1051–1054

Svanes T, Martinius AW, Hegre J, Maret J-P, Mjos R, Molina JCU (2004) Integration of subsurface applications to develop a dynamic stochastic modeling workflow. Am Assoc Petr Geol B 88:1369–1390

Szerbiak RB, McMechan GA, Corbeanu R, Forster C, Snelgrove SH (2001) 3-D characterization of a clastic reservoir analogue: from 3-D GPR data to a 3-D fluid permeability model. Geophysics 66:1026–1037

Tabor NJ, Poulsen CJ (2008) Paleoclimates across the Late Pennyslvanian-Early Permian tropical palaeolatitudes: A review of climate indicators, their distribution, and relation to

palaeophysiographic climate factors. Palaeogeography, Palaeoclimatology, Palaeoecology 268:293–310
Takano O, Waseda A (2003) Sequence stratigraphic architecture of a differentially subsiding bay to fluvial basin: the Eocene Ishikari Group, Ishikari Coal Field, Hokkaido, Japan. Sediment Geol 160:131–158
Tandon SK, Gibling MR, Sinha R, Singh V, Ghazanfari P, Dasgupta AS, Jain M, Jain V (2006) Alluvial valleys of the Gangetic Plains, India: timing and causes of incision. In: Dalrymple RW, Leckie DA, Tillman RW (eds) Incised valleys in time and space, vol. 85. Society for Sedimentary Geology (SEPM), Special Publication, pp 15–35
Tandon SK, Sinha R (2007) Geology of large river systems. In: Gupta A (ed) Large rivers: geomorphology and management. John Wiley and Sons, Chichester, pp 7–28
Tapponnier P, Peltzer G, Armijo R (1986) On the mechanics of the collision between India and Asia. In: Coward MP, Ries AC (eds) Collision tectonics, vol. 19. Geological Society of London, Special Publication, pp 115–157
Taylor CFH (1999) The role of overbank flow in governing the form of an anabranching river: the Fitzroy River, northwestern Australia. In: Smith ND, Rogers J (eds) Fluvial Sedimentology VI, vol. 28. Special Publication of the International Association of Sedimentologists, pp 77–92
Thakur GC (1991) Waterflood surveillance techniques—a reservoir management approach. J Petrol Technol 43:1180–1192
Thakur GC, Satter A (1998) Integrated waterflood asset management. Pennwell, Tulsa, Oklahoma
Thomas RG, Smith DG, Wood JM, Visser J, Calverley-Range EA, Koster EH (1987) Inclined heterolithic stratification—terminology, description, interpretation and significance. Sediment Geol 53:123–179
Thorne CR, Russell APG, Alam MK (1993) Planform pattern and channel evolution of the Brahmaputra River, Bangladesh. In: Best JL, Bristow CS (eds) Braided rivers, vol. 75. Geological Society, London, Special Publication, pp 257–276
Törnqvist TE (1993) Holocene alternation of meandering and anastomosing fluvial systems in the Rhine-Meuse delta (central Netherlands) controlled by sea-level rise and subsoil erodibility. J Sediment Petrol 63:683–693
Törnqvist TE (1994) Middle and late Holocene avulsion history of the River Rhine (Rhine-Meuse delta, Netherlands). Geology 22:711–714
Tye RS (1991) Fluvial-sandstone reservoirs of the Travis Peak Formation, East Texas Basin. In: Miall AD, Tyler N (eds) The three-dimensional facies architecture of terrigenous clastic sediments, and its implications for hydrocarbon discovery and recovery, vol. 3. Society of Economic Paleontologists and Mineralogists Concepts and Models Series, pp 172–188
Tye RS (2004) Geomorphology: an approach to determining subsurface reservoir dimensions. Am Assoc Petr Geol B 88:1123–1147
Tye RS, Bhattacharya JP, Lorsong JA, Sinfdelar ST, Knowck DG, Puls DD, Levinson RA (1999) Geology and stratigraphy of fluvio-deltaic deposits in the Ivishak formation: applications for development of Prudhoe Bay field, Alaska. Am Assoc Petr Geol B 83:1588–1623
Tyler N, Finley RJ (1991) Architectural controls on the recovery of hydrocarbons from sandstone reservoirs. In: Miall AD, Tyler N (eds) The three-dimensional facies architecture of terrigenous clastic sediments, and its implications for hydrocarbon discovery and recovery, vol. 3. Society of Economic Paleontologists and Mineralogists, Concepts in Sedimentology and Paleontology, pp 1–5
Tyler N, Galloway WE, Garrett CM Jr, Ewing TE (1984) Oil accumulation, production characteristics, and targets for additional recovery in major oil reservoirs of Texas. The University of Texas, Bureau of Economic Geology, Geological Circular 84-2, 31 p
Vail PR, Mitchum RM Jr., Todd RG, Widmier JM, Thompson III S,, Sangree JB, Bubb JN, Hatlelid WG (1977) Seismic stratigraphy and global changes of sea-level. In: Payton CE (ed) Seismic stratigraphy—applications to hydrocarbon exploration, vol. 26. American Association of Petroleum Geologists Memoir, pp 49–212

Vakarelov BK, Bhattacharya JP, Nebrigic DD (2006) Importance of high-frequency tectonic sequences during greenhouse times of earth history. Geology 34:797–800

Vandenberghe J (1993) Changing fluvial processes under changing periglacial conditions. Z Geomorph N.F. 88:17–28

Vandenberghe J, Kasse C, Bohnke S, Kozarski S (1994) Climate-related river activity at the Weichselian-Holocene transition: a comparative study of the Warta and Maas rivers. Terra Nova 6:476–485

Van der Zwan CJ (2002) The impact of Milankovitch-scale climatic forcing on sediment supply. Sediment Geol 147:271–294

Van Wagoner JC, Mitchum RM, Campion KM, Rahmanian VD (1990) Siliciclastic sequence stratigraphy in well logs, cores, and outcrops. American Association of Petroleum Geologists Methods in Exploration Series, vol. 7, 55 p

Varban BL, Plint AG (2008) Sequence stacking patterns in the Western Canada foredeep: influence of tectonics, sediment loading and eustasy on deposition of the Upper Cretaceous Kaskapau and Cardium formations. Sedimentology 55:395–421

Villalba M, Mendez O, Marcano C (2001) Opportunities for redevelopment of mature fields by determinations of hydraulic units for commingled production. Society of Petroleum Engineers, paper SPE 69599

Walker RG (1976) Facies models 1. General introduction. Geosci Canada 3:21–24

Walker RG (1990) Perspective—facies modeling and sequence stratigraphy. J Sediment Petrol 60:777–786

Wang Y, Straub KM, Hajek EA (2011) Scale-dependent compensational stacking: an estimate of autogenic time scales in channelized sedimentary deposits. Geology 39:811–814

Wanless HR (1964) Local and regional factors in Pennsylvanian cyclic sedimentation. In: Merriam DF (ed) Symposium on cyclic sedimentation, vol. 169. Kansas Geological Survey Bulletin, pp 593–605

Weber KJ, Van Geuns LC (1990) Framework for constructing clastic reservoir simulation models. J Petrol Technol 42:1248–1253, 1296–1297

Weissman GS, Hartley AJ, Nichols GJ, Scuderi LA, Olson ME, Buehler HA, Banteah R (2010) Fluvial form in modern continental sedimentary basins: distributive fluvial systems (DFS). Geology 38:39–42

Weissman GS, Hartley AJ, Nichols GJ, Scuderi LA, Olson ME, Buehler HA, Massengill LC (2011) Alluvial facies distributions in continental sedimentary basins—distributive fluvial systems. In: Davidson SK, Leleu S, North CP (eds) From river to rock record, vol. 97. Society for Sedimentary Geology (SEPM), Special Publication, pp 327–355

Wellner RW, Bartek LR (2003) The effect of sea level, climate, and shelf physiography on the development of incised-valley complexes: a modern example from the East China Sea. J Sediment Res 73:926–940

Wells NA, Dorr JA (1987) Shifting of the Kosi River, northern India. Geology 15:204–207

Wescott WA (1993) Geomorphic thresholds and complex response of fluvial systems—some implications for sequence stratigraphy. Am Assoc Petr Geol B 77:1208–1218

Westrich JT, Fuex A, O'Neal PM, Halpern HI (1999) Evaluating reservoir architecture in the northern Gulf of Mexico with oil and gas chemistry, vol. 2. SPE Reservoir Evaluation and Engineering, pp 514–519 (SPE paper 59518)

Willis AJ (2000) Tectonic control of nested sequence architecture in the Sego Sandstone, Neslen formation, and Upper Castlegate Sandstone (Upper Cretaceous), Sevier Foreland Basin, Utah, USA. Sediment Geol 136:277–317

Willis BJ, White CD (2000) Quantitative outcrop data for flow simulation. J Sediment Res 70:788–802

Wood JM, Hopkins JC (1992) Traps associated with paleovalleys and interfluves in an unconformity bounded sequence: lower Cretaceous Glauconitic Member, southern Alberta, Canada. Am Assoc Petr Geol B 76:904–926

Wood L (2007) Quantitative seismic geomorphology of Pliocene and Miocene fluvial systems in the northern Gulf of Mexico, USA. J Sediment Res. 77:713–730

Wright AM, Ratcliffe KT, Zaitlin BA, Wray DS (2010) The application of chemostratigraphic techniques to distinguish compound incised valleys in low-accommodation incised-valley systems in a foreland-basin setting: an example from the Lower Cretaceous Manville Group and Basal Colorado Sandstone (Colorado Group), Western Canada Sedimentary Basin. In: Ratcliffe KT, Zaitlin BA (eds) Application of modern stratigraphic techniques: theory and case histories, vol. 94. Society for Sedimentary Geology (SEPM), Special Publication, pp 93–107

Wright VP, Marriott SB (1993) The sequence stratigraphy of fluvial depositional systems: the role of floodplain sediment storage. Sediment Geol 86:203–210

Yalin MS (1992) River mechanics. Pergamon Press, Oxford, 219 p

Yu X, Ma X, Qing H (2002) Sedimentology and reservoir characteristics of a Middle Jurassic fluvial system, Datong basin, northern China. B Can Petrol Geol 50:105–117

Yangquan J, Jiaxin Y, Sitian L, Ruiqi Y, Fengjiang L, Shengke Y (2005) Architectural units and heterogeneity of channel reservoirs in the Karamay Formation, outcrop area of Karamay oil field, Junggar basin, northwest China. Am Assoc Petr Geol B 89:529–545

Yoshida S (2000) Sequence stratigraphy and facies architecture of the upper Blackhawk Formation and the Lower Castlegate Sandstone (Upper Cretaceous), Book Cliffs, Utah, USA. Sediment Geol 136:239–276

Yoshida Shuji, Willis A, Miall AD (1996) Tectonic control of nested sequence architecture in the Castlegate Sandstone (Upper Cretaceous), Book Cliffs, Utah. J Sediment Res 66:737–748

Zaitlin BA, Warren MJ, Potocki D, Rosenthal L, Boyd R (2002) Depositional styles in a low accommodation foreland basin setting: an example from the Basal Quartz (Lower Cretaceous), southern Alberta. B Can Petrol Geol 50:31–72

Zeng H (2007) Seismic imaging for seismic geomorphology beyond the seabed: potential and challenges. In: Davies RJ, Posamentier HW, Wood LJ, Cartwright JA (eds) Seismic geomorphology: applications to hydrocarbon exploration and production, vol. 277. Geological Society, London, Special Publication, pp 15–28

Zeng HL, Hentz TF (2004) High-frequency sequence stratigraphy from seismic sedimentology: applied to Miocene, Vermilion Block 50, Tiger Shoal area, offshore Louisiana. Am Assoc Petr Geol B 88:153–174

Zeng X, McMechan GA, Bhattacharya JP, Aiken CLV, Xu X, Hammon WS III, Corbeanu RM (2004) 3-D imaging of a reservoir analogue in point bar deposits in the Ferron Sandstone, Utah, using ground-penetrating radar. Geophys Prospect 52:151–163